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ON THE FORMATION OF HURRICANE ALICE, 1955

With Notes on Other Cold-Season Tropical Storms

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ABSTRACT

The occurrence of hurricane Alice in January 1955 was a meteorologically unseasonable event. The synoptic history of the storm during its genesis over the warm waters of the tropical Atlantic, its intensification while drifting west-southwestward, its passage over the Leeward Islands, and its decay in the eastern Caribbean is reviewed. Its formation is attributed to a rare combination of favorable circumstances. These included anticyclogenesis and blocking in the middle latitudes of the western Atlantic and strong, deep easterlies to the south. This prolonged pattern of circulation effectively isolated the tropical regions from polar air invasions and caused an incipient cyclone that had formed at the point of fracture of an extended Atlantic trough to move west-southwestward over relatively warm waters where it was transformed into a warm-core storm of hurricane intensity. The wind and thermal structure is analyzed and land observations and damage estimates are presented. Finally, some comparative notes on other cold-season storms in the Tropics are given.

1. INTRODUCTION

Hurricane Alice of 1955 gave rise to considerable speculation at the time of its formation due to its unseasonable occurrence. This storm had its genesis in a low pressure center which formed during December 29, 1954, around latitude 21° N., longitude 49° W. (fig. 1). It then drifted west-southwestward, gradually developing warm-core properties, and intensified to hurricane proportions by the afternoon of January 1, 1955. During January 2 the storm passed over the Leeward Islands causing considerable damage.

The formation of storms of this type in the Atlantic Ocean in winter is so infrequent that the events leading up to their formation must constitute a coincidence of favorable circumstances very seldom duplicated. The main purpose of this report is to describe those conditions that led to the formation of Alice. In addition, a discussion of the wind and thermal structure of the storm is given and some remarks on other cold-season tropical storms are made.

The analysis of Alice is based on the surface maps, the low-level wind charts, contour charts at the standard isobaric levels, and time cross-sections at key stations. Low-level wind maps were obtained by combining ship wind reports and Pibal winds at 2,000 feet at land stations. This procedure for the combination of wind data has been used previously by the writer with good results. Upper-air charts at 850, 700, 500, and 200-mb. levels were available, but only the 500-mb. charts are illustrated here. The analysis over the ocean is mainly dependent on ship reports and the best coverage is usually available during the daytime. As a consequence, the surface analysis is presented for 1230 GMT and the upper-air maps for 1500 GMT, although for upper-air analysis in these regions the 0300 GMT observations are generally recommended.

The upper-air analysis over the ocean is based to a large extent on extrapolations from ship observations. Extrapolations were made first from the surface to 700 mb. by estimating the 700-mb. temperature and assuming that the mean temperature of the layer was equal to the

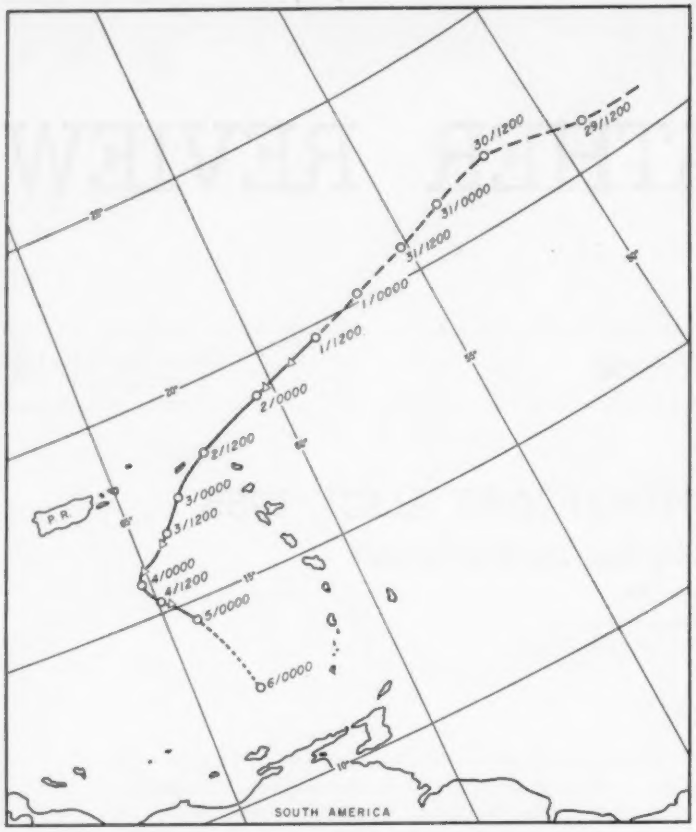

FIGURE 1.—Track of hurricane Alice, December 29, 1954–January 5, 1955. Dates and times (GMT) are indicated.

arithmetic average of the surface and 700-mb. temperatures. The 700-mb. chart was then analyzed and necessary adjustments made. The 500-mb. heights were extrapolated from the 700-mb. level using the same procedure.

2. SYNOPTIC HISTORY

The beginnings of what eventually led to the formation of Alice started around December 25–26, 1954, from a typical wintertime picture. The charts for December 26 (fig. 2) show a frontal system extending southwestward from an extratropical Low (centered around 45° N., 45° W.) and becoming an E-W quasi-stationary front in the region of the Bahama Islands. This front had moved rapidly southeastward from the continent in the previous 48 hours and passed Bermuda around 1230 GMT, December 25. On the 26th a pertubation existed in the trade easterlies around longitude 53°–54° W. extending southward from the frontal trough in the north. This type of formation, which is frequent in this area during the cold season, has been described in the literature [1]. Although the wind structure is similar to that of the summertime easterly waves, these cold-season perturbations form in connection with a polar system and move to the east against the trade winds. One other system that predominated on the 26th was the anticyclone behind the cold front (fig. 2). The behavior of this high cell was of particular significance in the events that followed. At this time the anticyclone had a well-defined NNE–SSW elongated axis centered along the coast.

At upper levels (fig. 2B) the flow was dominated by the extended polar trough associated with the surface system. The extension of the trough into the Tropics was clearly defined by the winds at San Juan, Guadeloupe, and Trinidad. To the west of the trough a well-defined ridge extended northward along the east coast.

During the next 2 days the polar system and associated tropical perturbation moved steadily eastward steered by the westerlies. On December 27 (fig. 3A) the extratropical Low was located around longitude 40° W. The tropical trough moved to around longitude 49° W.—an eastward speed of around 12 knots. As there were not many data near the position of the trough, the analysis indicated on this and other charts was based on consideration of the intermediate maps, on the analysis of the low-level wind charts, and also on proper continuity.

Significant changes took place in the shape of the anticyclone. The center of the cell moved very slowly south-southeastward and the major axis turned to a more E-W orientation. The polar front moved slightly southward in the region north of Cuba and Hispaniola. The temperature discontinuity was not too clear, but there was a wind speed discontinuity with higher speeds north of the front. Showers were reported along the frontal zone in the eastern Bahama Islands. At the 500-mb. level (fig. 3B) the polar trough moved eastward and passed Ship Echo (35° N., 48° W.). At the same time a sharp ridge spread into the region north of Bermuda. Contour rises were observed at Bermuda and particularly at Sable Island. The Bermuda winds became more northerly; ordinarily they should turn to westerly with the normal approach of a ridge. A secondary trough formed between Bermuda and San Juan in an ENE-WSW orientation.

By December 28 (fig. 4A) the transformation of the High into a well-developed E-W oriented anticyclone was completed. This anticyclone dominated the flow from longitude 40° W. westward to the Eastern States and easterly flow prevailed south of latitude 35° N. The polar Low and tropical wave continued moving eastward at about the same speed. The actual position of the southern section of the tropical perturbation was uncertain, but its northern portion was well delineated by a series of ship observations around latitudes 22°–27° N. and longitudes 36°–46° W. The large amplitude of the 1016-mb. isobar suggests the presence of a small low center. Considerable cloudiness and weather were observed over this area.

Along the quasi-stationary front no temperature discontinuity was discernible any more, except in the region north of latitude 25° N. A wind speed discontinuity was still present, but as it generally occurs in this area the polar front was gradually losing its properties.

The low-level wind map (dashed pattern in fig. 4B) shows the frontal trough and tropical wave as an extended

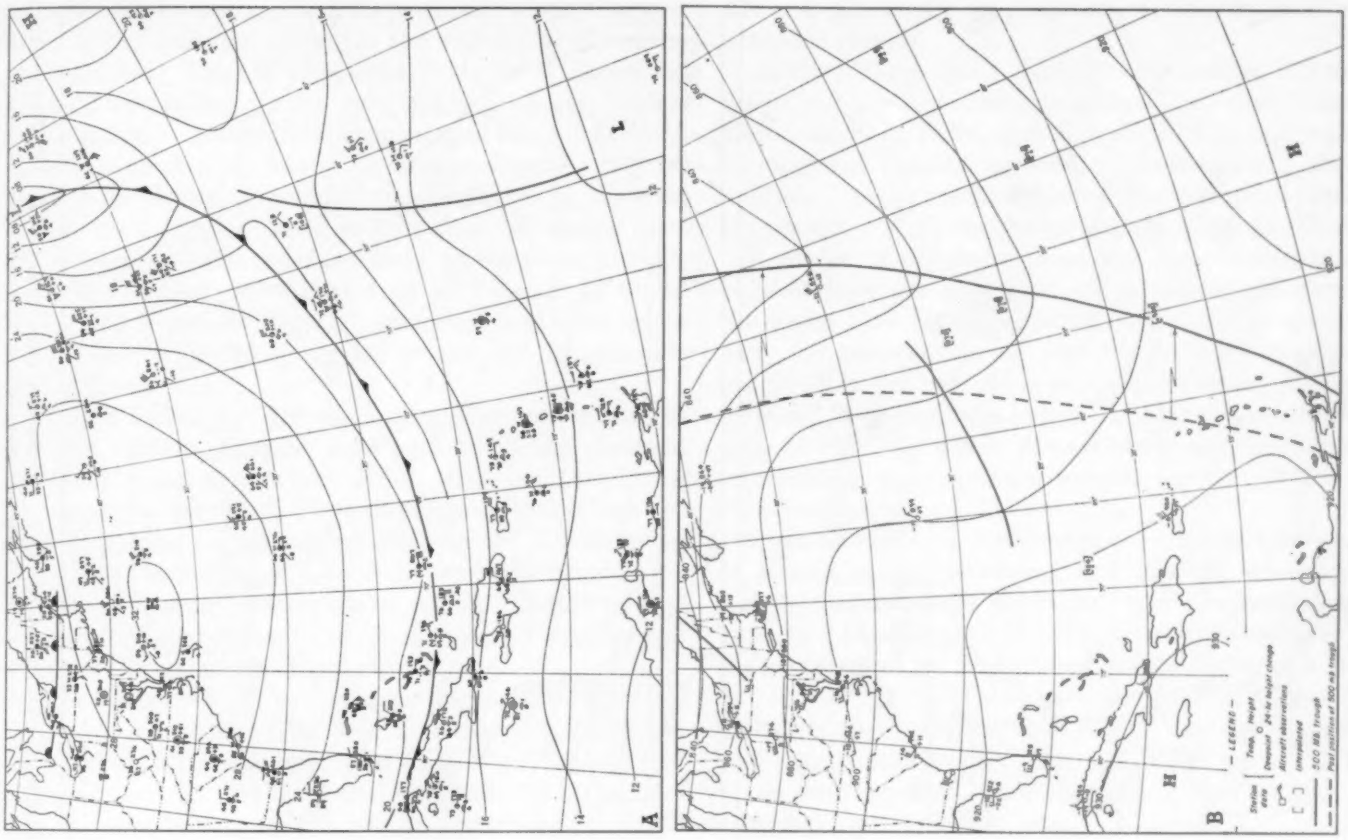

FIGURE 3.—December 27, 1954. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

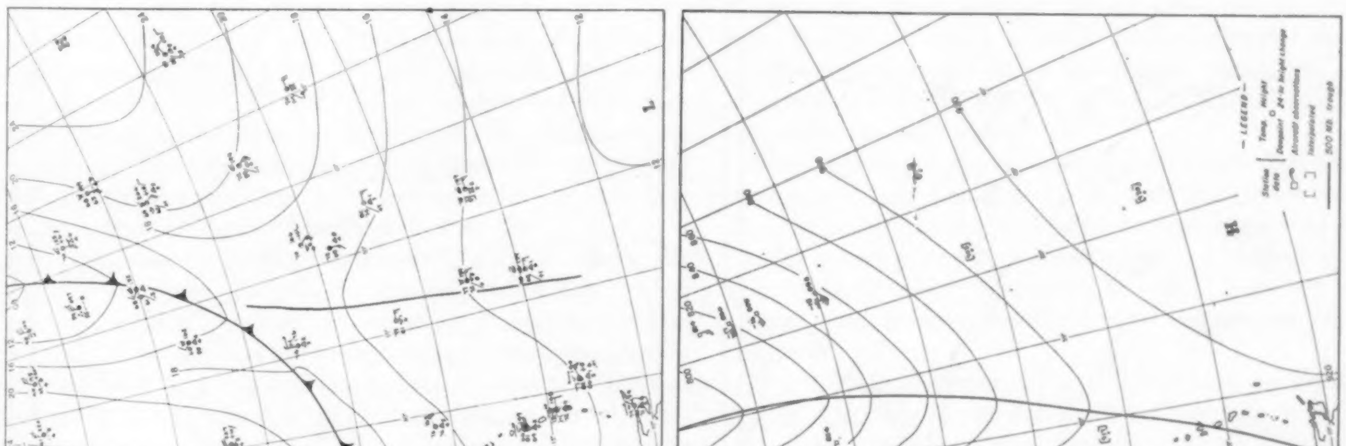

FIGURE 2.—December 26, 1954. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

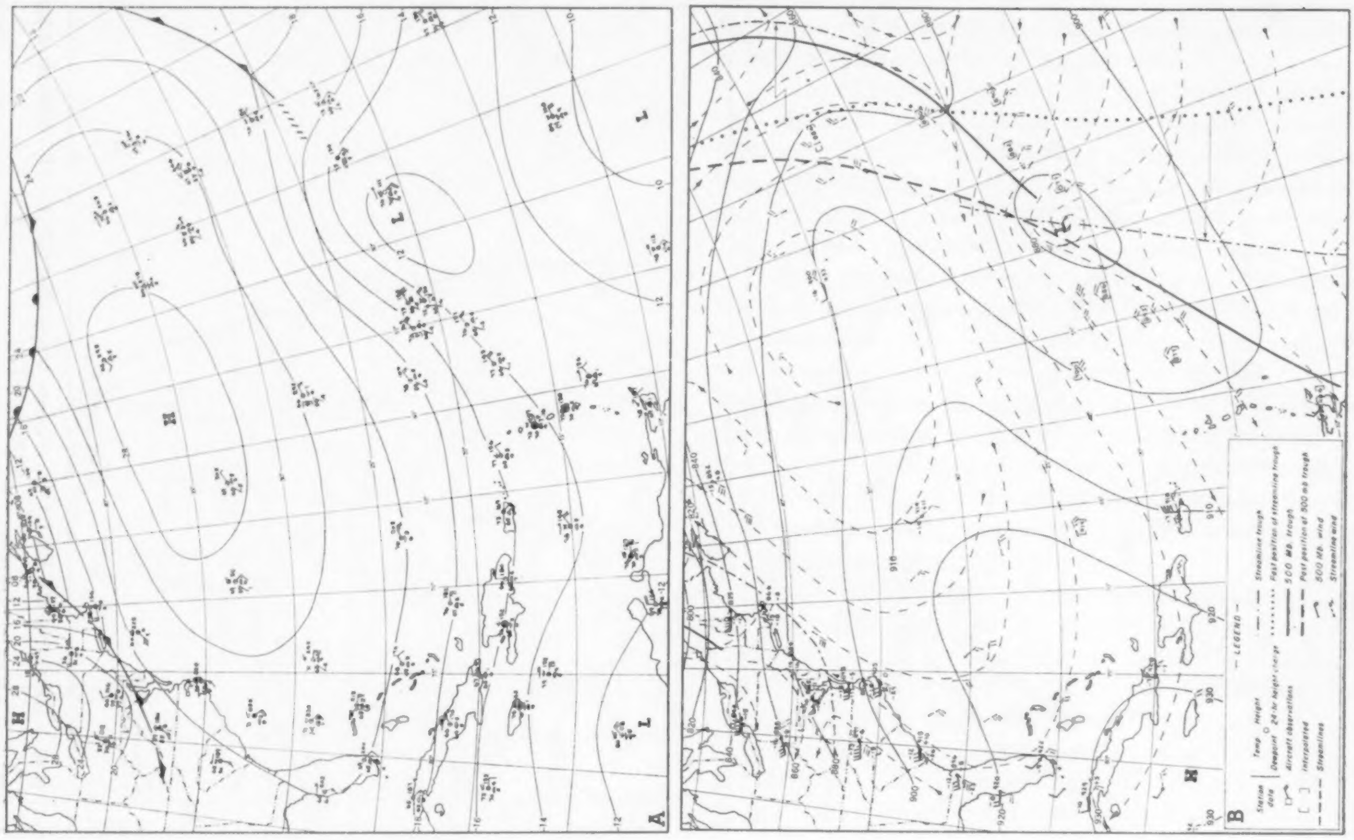

FIGURE 5.—December 29, 1954. (A) Surface map, 1230 GMT. (B) 500-mb. map with 2,000-foot streamlines, 1500 GMT.

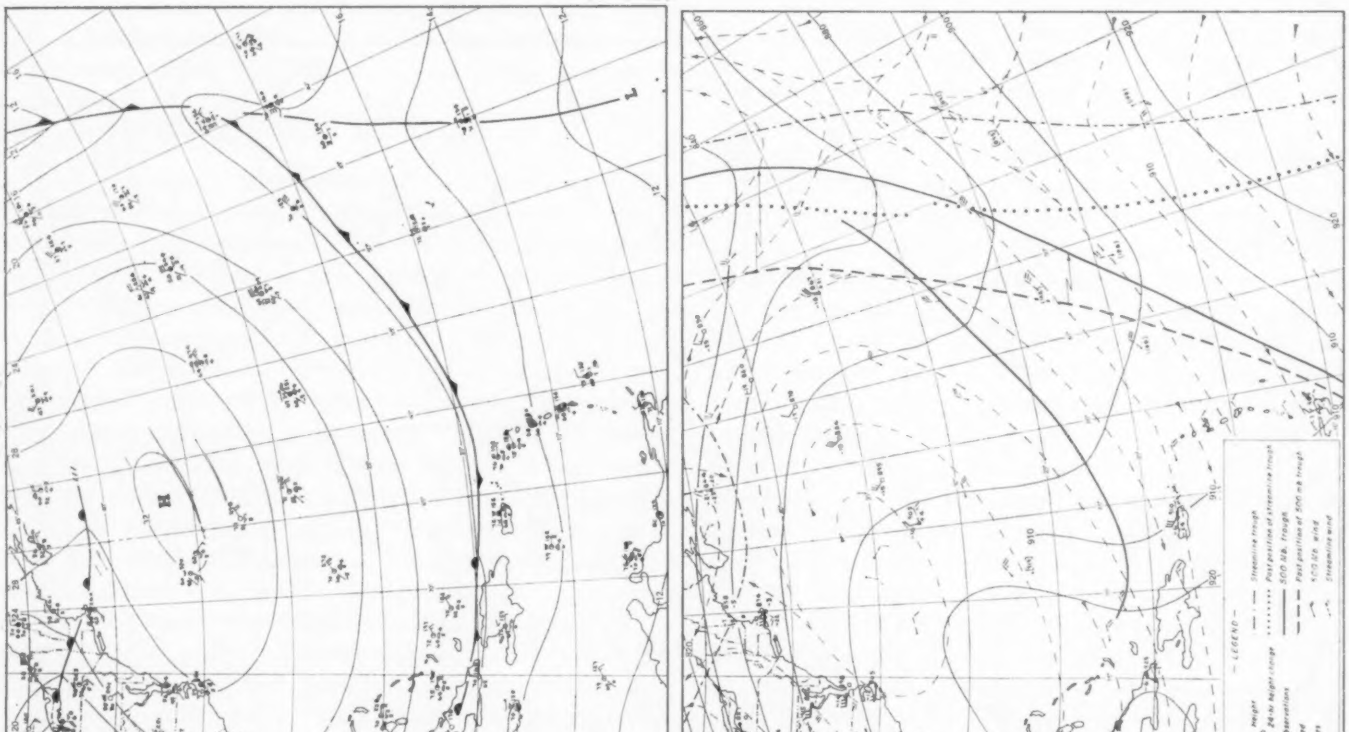

FIGURE 4.—December 28, 1954. (A) Surface map, 1230 GMT. (B) 500-mb. map with 2,000-foot streamlines, 1500 GMT.

trough. There was no wind discontinuity along the polar front in the Atlantic, except in the vicinity of the extra-tropical Low. The SW wind at 24.5° N., 40° W. suggested a closed circulation in the northern end of the tropical perturbation. After careful inspection of this wind report, there was no obvious reason for disregarding it. It gives the first indication of the ensuing cyclogenesis in this area.

Over the Caribbean area, a fairly deep and strong easterly current prevailed at this time. At San Juan (fig. 12) the east flow was 12,000 feet deep with speeds of 20–30 knots. At Bermuda (fig. 11) easterlies prevailed up to 25,000 feet, while 48 hours before no east components were present.

At high levels (fig. 4B) the anticyclogenesis continued over the western Atlantic; rises were still being observed at Bermuda and also at Ship Echo. The main trough in the westerlies continued its eastward progression, but the tropical section stayed behind close to the Antilles. In the belt of the westerlies the strong current persisted over the New England-Newfoundland region. Rapid motion of weak migratory troughs in these strong westerlies was observed throughout this period.

After December 28, cyclogenesis began in the tropical Atlantic. Already at 1230 GMT, December 28 (fig. 4A) a closed circulation was indicated in the junction of the extended trough. By 0030 GMT, December 29, the fracture of the extended trough was completed and a closed Low formed in the northern portion of the tropical wave. This tropical Low then came under the influence of the anticyclone and reversed its motion to the west. The polar section of the trough continued its steady motion to the east in the belt of the westerlies. At 1230 GMT, December 29 (fig. 5) the closed Low was centered at approximately 21° N., 49° W., already embedded in the easterly current. This position gave a westward motion of around 14 knots from the position of the wave 24 hours earlier; this suggests that the chart of December 28 marked the easternmost position of the extended trough and that fracture and cyclogenesis started at that time.

A comparison of figure 5B and figure 4B shows the changes that took place in the low-level wind flow during the process of fracture of the extended trough and cyclogenesis. The position of the disturbance was considerably farther south and west of the incipient closed circulation indicated on the previous day. Such things occur sometimes during the first stages of cyclogenesis; the process is probably dissolution of one center and reformation in another section of the wave rather than continuous motion of the same circulation.

No fronts have been analyzed in this Low. There was no evidence of a frontal wave in the process of formation. Cyclogenesis seems to have occurred by a process similar to that observed sometimes during the summer. During the summer, extended troughs are usually formed by the junction of an easterly wave travelling westward and a trough in the westerlies moving eastward. Cyclogenesis in the easterly current in the manner indicated by figures 4B and 5B occurs on occasions during fracture of the extended trough.

At the 500-mb. level (fig. 5B) the analysis is based almost entirely on extrapolations. The observations at Bermuda, Ship Echo, and San Juan, together with the extrapolated heights, provide a consistent and plausible analysis. An upper Low formed in the tropical section of the trough. With the continued extension to the northeast of the anticyclonic nose, this Low was practically isolated from the westerlies. The indications were that the upper Low formed simultaneously and in connection with the surface Low as part of the same cyclogenetic process. In the belt of the westerlies a strong jet current persisted; a trough was present in Canada around longitude 65°–70° W., which was evidently associated with an intensifying wave cyclone present at the surface over Newfoundland.

From this time on the story of the tropical Low was one of steady motion westward and gradual warming and transformation into a warm-core type of circulation. At 1230 GMT, December 30 (fig. 6A) three ship observations in the periphery of the Low pinpointed the circulation around 22° N., 52° W. and also gave the first indications of a closed circulation in the wind field. The central pressure was approximately 1008 mb. and wind speeds of 30–35 knots were reported. Cloudiness and bad weather were observed over the entire area.

In the belt of the westerlies intense cyclogenesis occurred. The frontal Low that was located the previous day over Newfoundland intensified considerably and moved rapidly eastward at a speed of around 37 knots. At the same time another frontal wave moved into the New York area. As a result of this cyclogenesis the anticyclone was squeezed into a flat two-cell structure. Nevertheless, it held its position and blocked the southward motion of the cold front.

The ridge line persisted also at high levels (fig. 6B) and the tropical Low was already isolated from the westerlies. The old trough in the westerlies moved away and a new rapidly moving trough associated with a surface cyclone was centered along longitude 45° W. Another rather intense system was observed over the region of Lake Ontario.

The surface analysis of the tropical Low on December 31 (fig. 7A) is based on continuity. No data were available at this time, but observations received 6 hours later indicated the Low, with central pressure of around 1007 mb., was moving westward at about 8 knots. Wind speeds of about 35–40 knots continued to be observed.

Strong cyclogenesis occurred once again in the belt of the westerlies. The previous cyclone continued the rapid motion and disappeared from the picture, but another occluded Low developed in the New England-Newfoundland area. The cold front advanced into central Florida and toward Bermuda, but it stayed far from the tropical Low. The ridge line was maintained E-W along latitude 32° N., although the high pressure cell was gradually

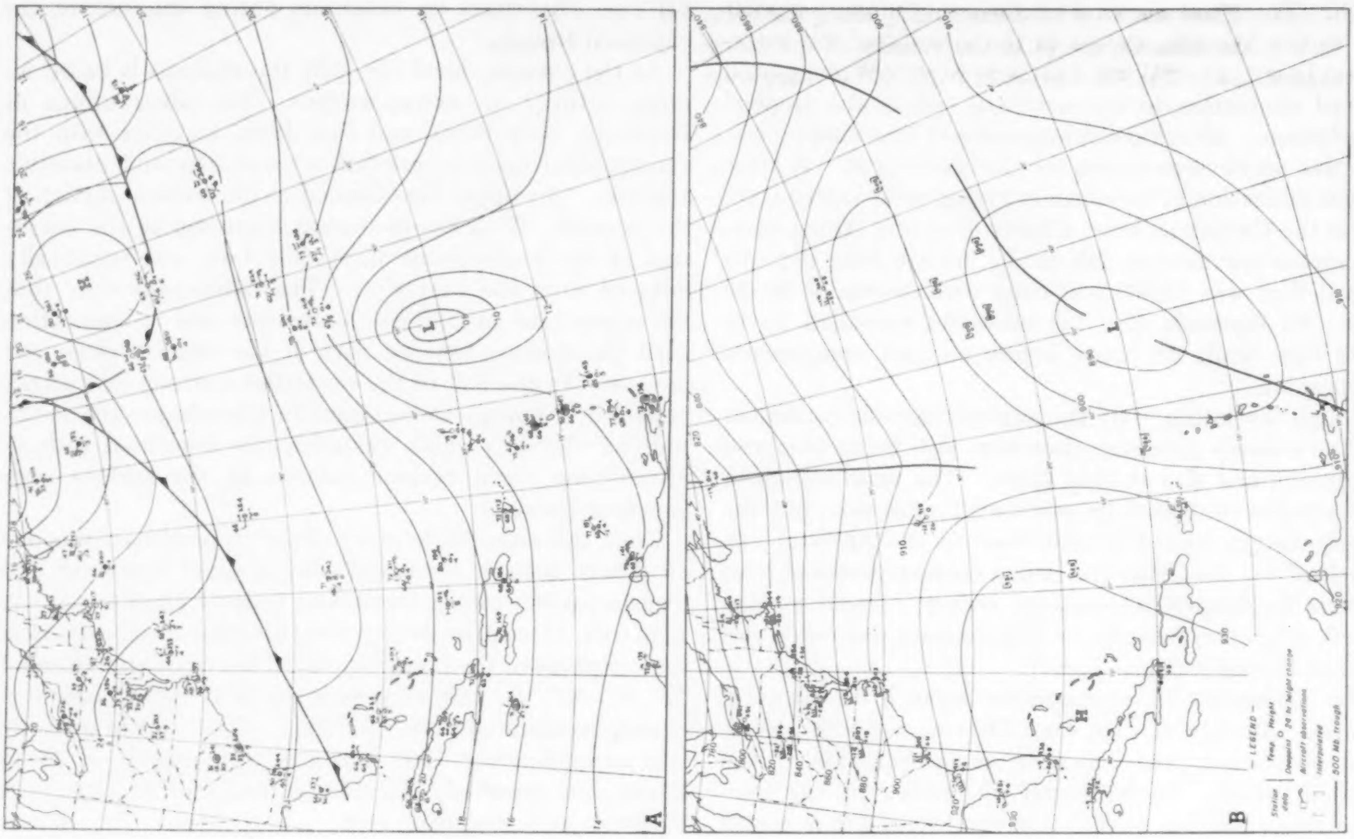

FIGURE 7.—December 31, 1954. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

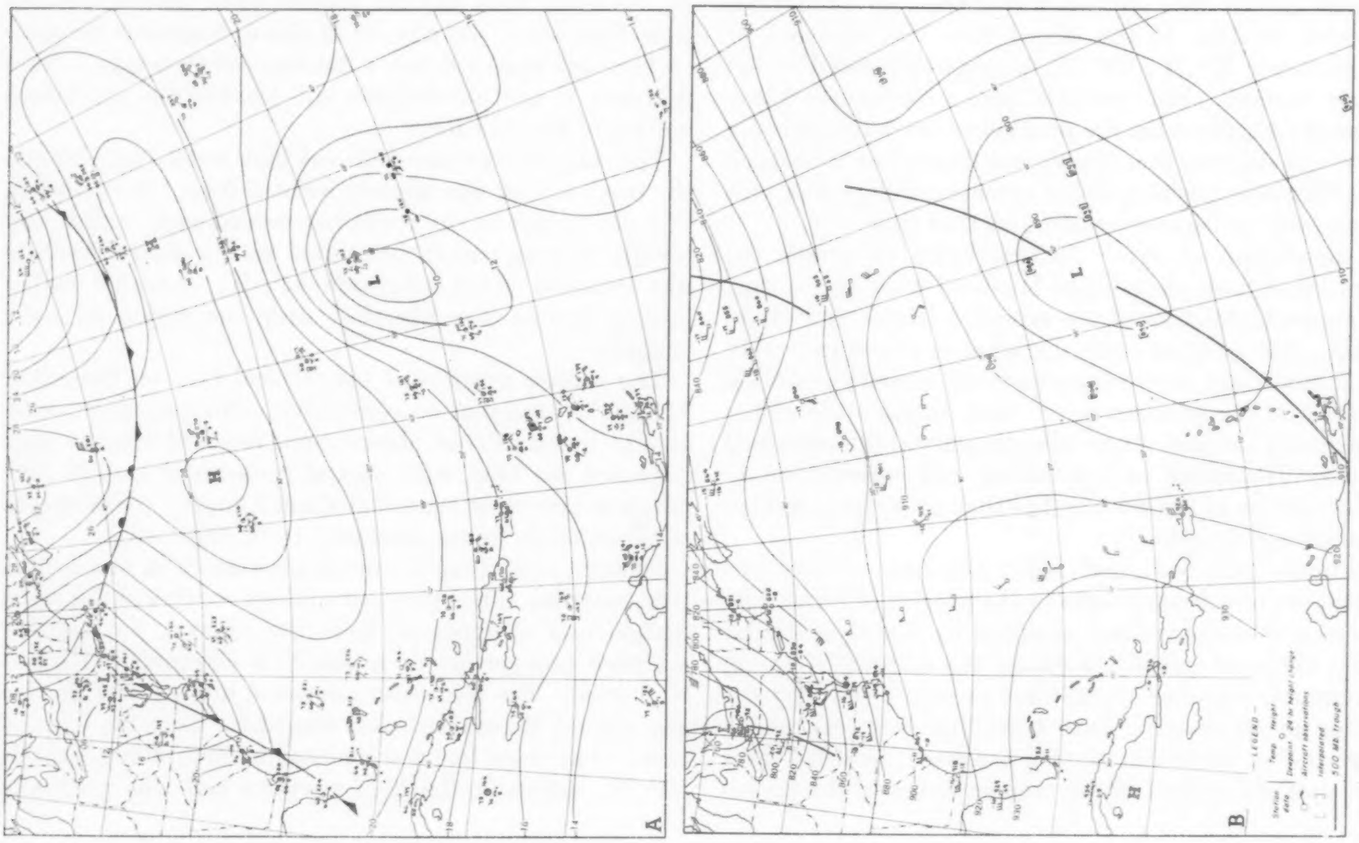

FIGURE 6.—December 30, 1954. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

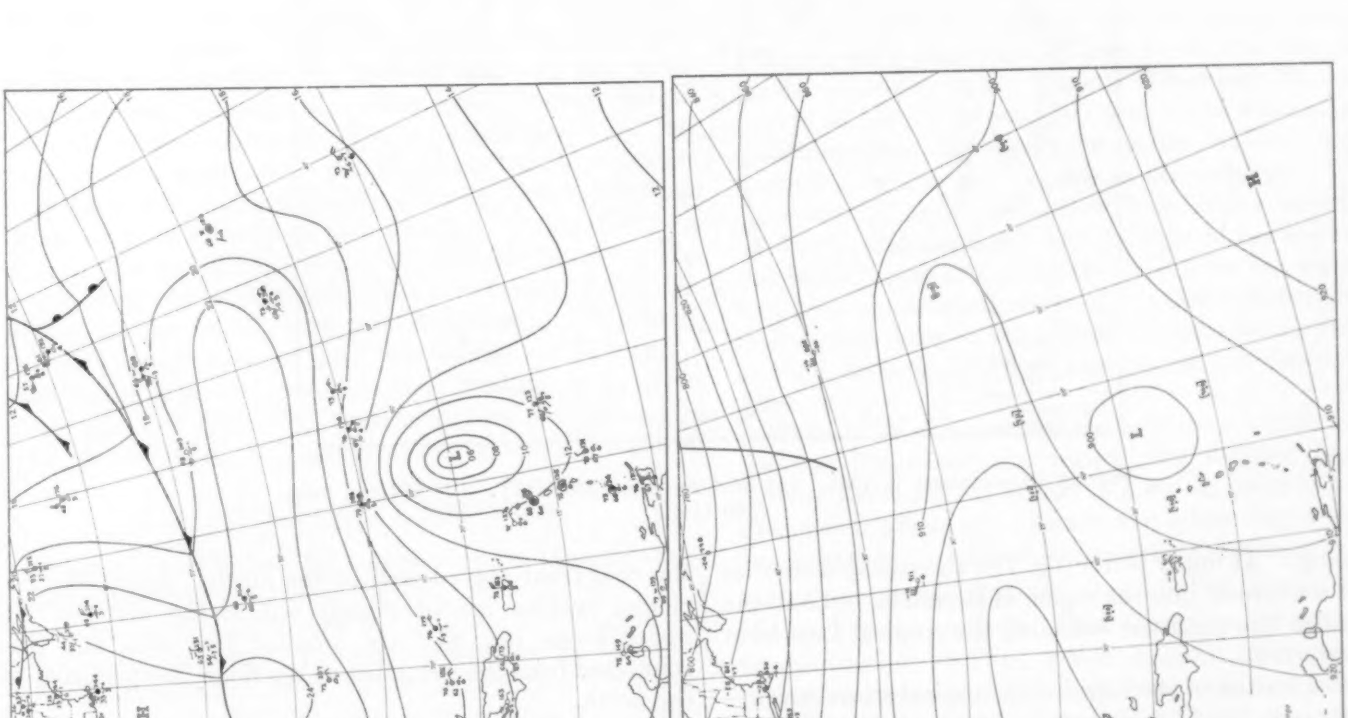

FIGURE 8.—January 1, 1955. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

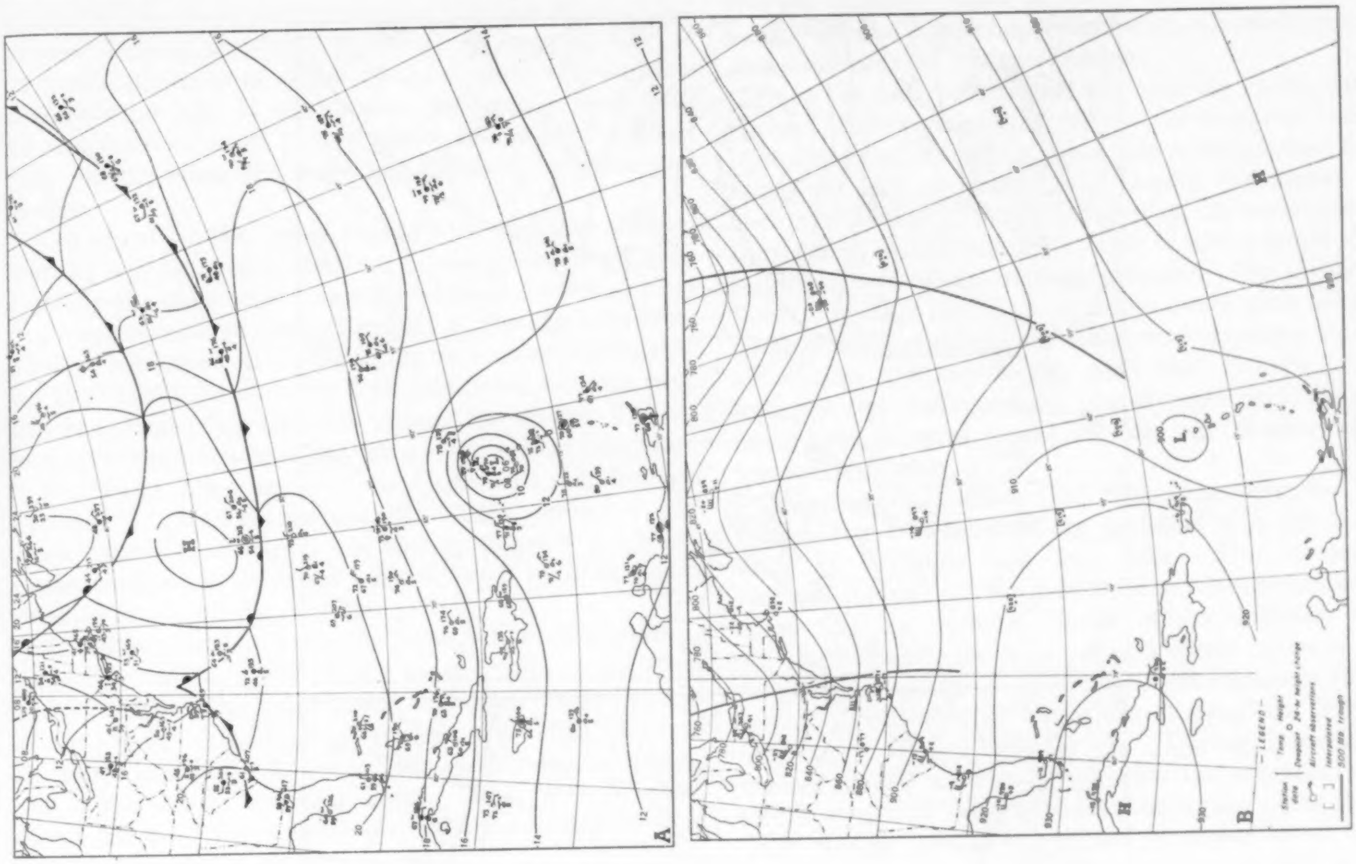

FIGURE 9.—January 2, 1955. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

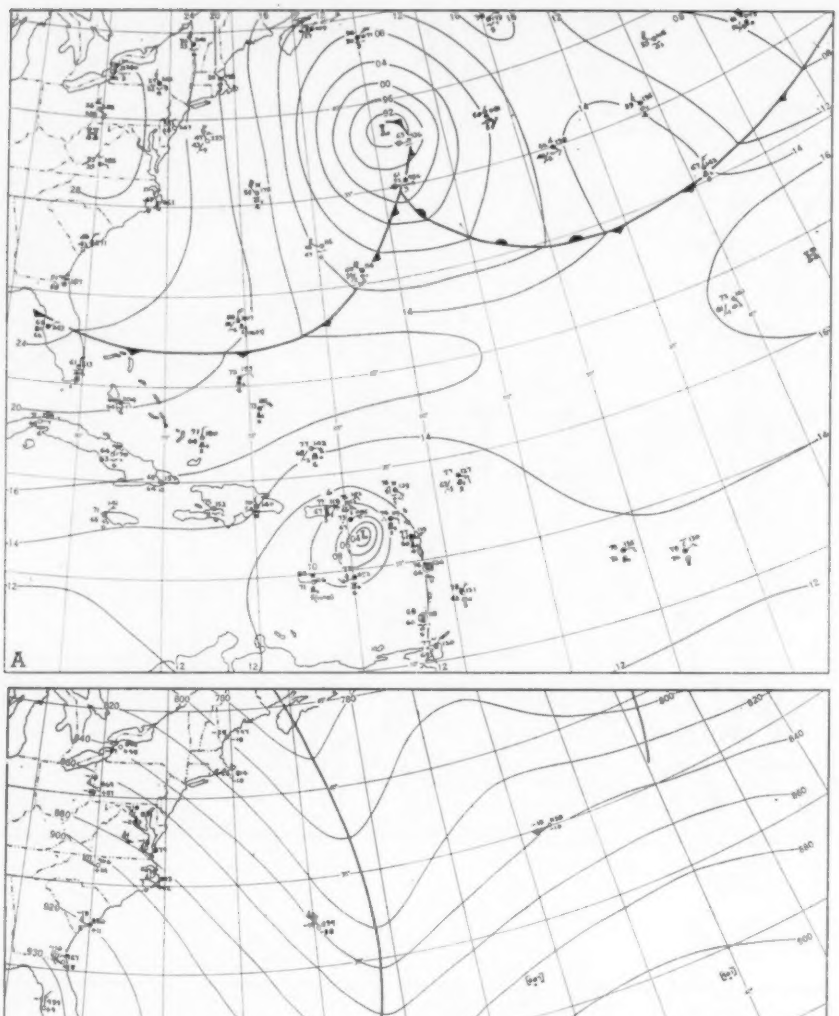

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Figure 10.—January 3, 1955. (A) Surface map, 1230 GMT. (B) 500-mb. map, 1500 GMT.

collapsing. At upper levels (fig. 7B) the strong westerlies spread southward into the region of Bermuda, but still the E-W ridge line persisted secluding the tropical Low from the westerlies.

Intensification of the Low into a tropical storm was apparently completed by January 1. The analysis of the Low indicated in figure 8A is based on continuity. A series of observations after this time indicated the presence of the tropical storm. As for developments in the westerlies, the high pressure belt was maintained north of the center and the cold front stayed well to the north. Again, the occluded cyclone moved rapidly out of the picture. At upper levels (fig. 8B) the tropical Low was still disconnected from the westerlies as the ridge line persisted to the north.

During the afternoon of January 1 a series of observations indicated the presence of a tropical storm circulation. The ship *Arawak* reported at 1919 GMT, January 1, at a position 19°15′ N., 59°10′ W., "West wind 12, barometer 987, temperature 66, visibility nil." Soon afterward a

United States Navy ship with radar facilities reported the center "located at 19°07′ N., 60°07′ W. moving 255° at 17 knots, diameter of center 20 miles, scattered showers, surface wind NW 47 knots, seas very rough, swell 040° 15 feet 6 seconds, sea level pressure 1000.1 mb." Other ship reports along the same line followed in quick succession.

The islands to the east of Puerto Rico were immediately warned against winds of storm intensity and high seas. Initially the circulation was referred to as a tropical Low, but reports received during January 2 from the land areas affected confirmed winds bordering on hurricane intensity. At 1900 GMT on January 2 an additional bulletin from the San Juan office officially named the system "Alice" and operations continued as for a summertime hurricane.

Hurricane Alice entered the Caribbean Sea late on January 2. Aircraft reconnaissance was made into the storm twice on January 3 and once on January 4. Maximum winds of 50–55 knots were reported by both flights into the storm on January 3.

The charts for January 2 (fig. 9) still show the persisting ridge line to the north of the storm. The polar front moved south of Bermuda but it still was outside the area of the storm. Much deeper troughs were then observed in middle latitudes. At the surface another frontal system was approaching the east coast. The 500-mb. chart shows one deep trough in the central Atlantic and another, also very intense, over the east coast.

On January 3 (fig. 10) the picture began to change radically. Strong cyclogenesis occurred in the Atlantic north of Bermuda. This new cyclone was in a more southern position than the previous ones and also much more intense. The ridge line was practically gone and the cold air moved southward into the Caribbean area. At high levels a deep trough moved to the east of Bermuda and the middle latitude troughs extended into the Tropics again after an interval of about 8 days. After this time, the circulation of Alice began to dissipate rapidly. The storm recurved to the south in the afternoon of January 3 (fig. 1) as the westerly flow was established over the area. Aircraft reconnaissance into the storm on January 4 reported only a wide area of squally weather with maximum winds of 35 knots. Soon afterward the circulation disappeared in the southeastern corner of the Caribbean Sea.

3. REMARKS ON FORMATION

As mentioned before, for a type of storm that occurs so infrequently the conditions for the formation must be a rare combination of favorable circumstances. Such was the case in this study. The main factor was the establishment and maintenance of the blocking High. This anticyclogenesis initiated and maintained a series of favorable conditions for the formation and intensification of the tropical Low. It maintained for a prolonged period a fairly strong, deep, and extensive easterly current. The spreading of the anticyclonic surge eastward over the western Atlantic ended in the isolation of the tropical section of the extended trough with the ensuing cyclogenesis during the fracture process.

Once the Low was formed and moving in the easterly current, the determining factor was the continued isolation from the polar air, which again was accomplished by the persistent high pressure belt. During the period there were two strong cyclones forming in the westerlies (see charts for Dec. 30 and 31) neither of which could extend its influence into the tropical latitudes. The presence of the strong High forced the cyclones into a path far to the north and the rapid motion in the westerly current did not allow time for interaction with the Tropics. Under ordinary circumstances any one of these cyclones would have been able to push the cold air southward into tropical latitudes.

The warming process that took place over the whole region after December 26 can be followed in the observations at Bermuda and San Juan. The temperature changes throughout the troposphere were determined by computing the 24-hour change in the thickness of the standard isobaric levels, which is a direct measure of the temperature change. The changes at Bermuda (fig. 11) indicated warming in the whole troposphere from December 27 until December 30. During December 31 cooling started at high levels with the extension of the strong westerlies southward into the region. (See fig. 7B.) This cooling spread downward to lower levels during January 1 and 2. At San Juan (fig. 12) cooling occurred at lower levels during December 28 and 29 with the advance southward of the E-W quasi-stationary cold front, while warming took place at high levels after the passage of the upper trough. During December 30 warming started at low levels probably due to the warmer air approaching from the Bermuda region—at this time the tropical Low was still far out in the Atlantic. Warmer air spread through the lower troposphere during December 31 and January 1 and 2. Part of the warm air in this later period was undoubtedly due to the warm current from the tropical storm. The new influx of cold air reached San Juan on January 3. This cooler air extended into the eastern Caribbean and started the dissipation of Alice.

While these temperature changes were taking place in the San Juan-Bermuda region, the tropical Low was moving westward embedded in a strong easterly current, disconnected from the cold air, and advancing over warm ocean waters. A total of 53 observations of water temperatures were available in the area 15°–30° N. and 40°–60° W. during the period. These gave an average water temperature of 76.6° F., which is about 1° warmer than the mean January water temperature for this area given in the Climatic Charts of the Oceans. Similarly, air temperatures for the same area averaged 74.6° F.; that is about 2° cooler. The air was thus advancing over warmer waters. At the same time the storm was isolated from

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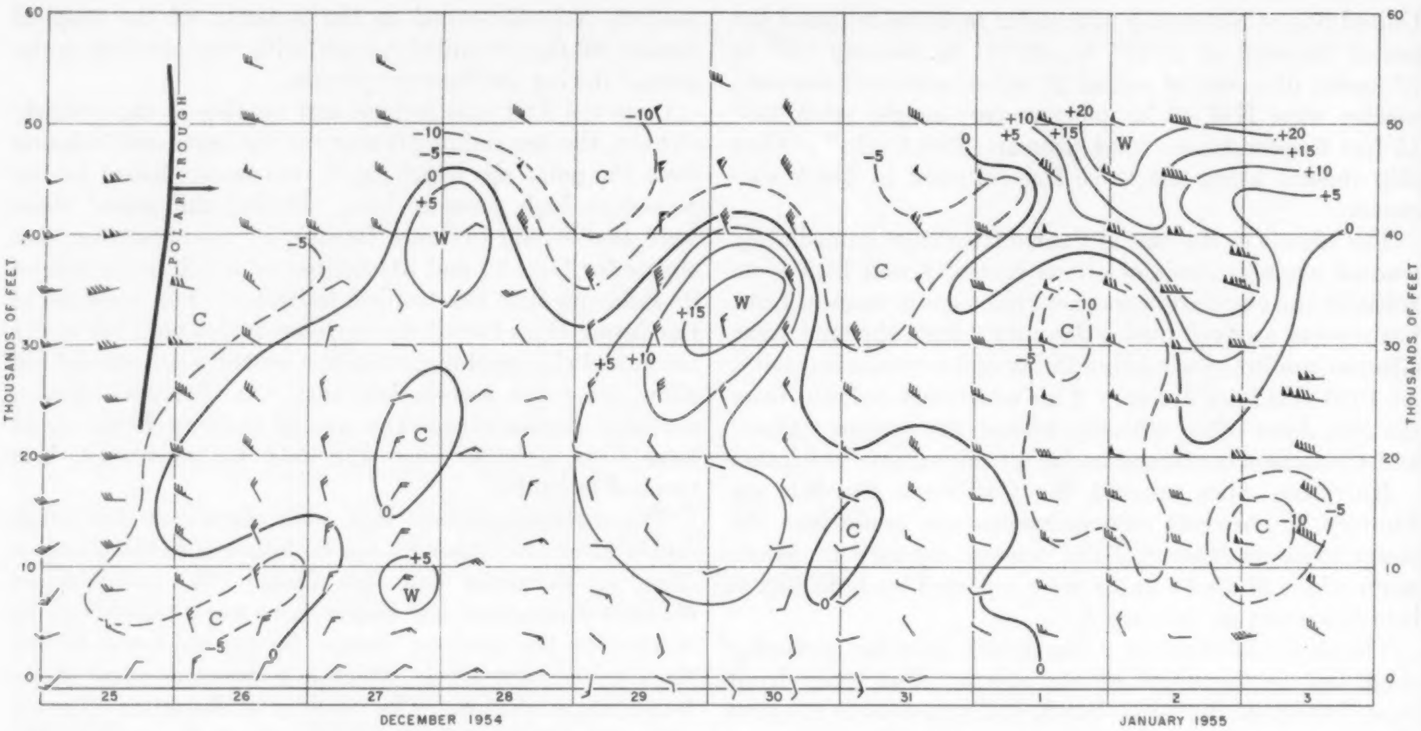

FIGURE 11.—Time section at Kindley Field, Bermuda, December 25, 1954–January 3, 1955, showing analysis of the 24-hour change in the thickness between the standard isobaric levels (in 10's of feet). W stands for warming, C, for cooling. Wind speeds in knots. Analysis displaced to center of period.

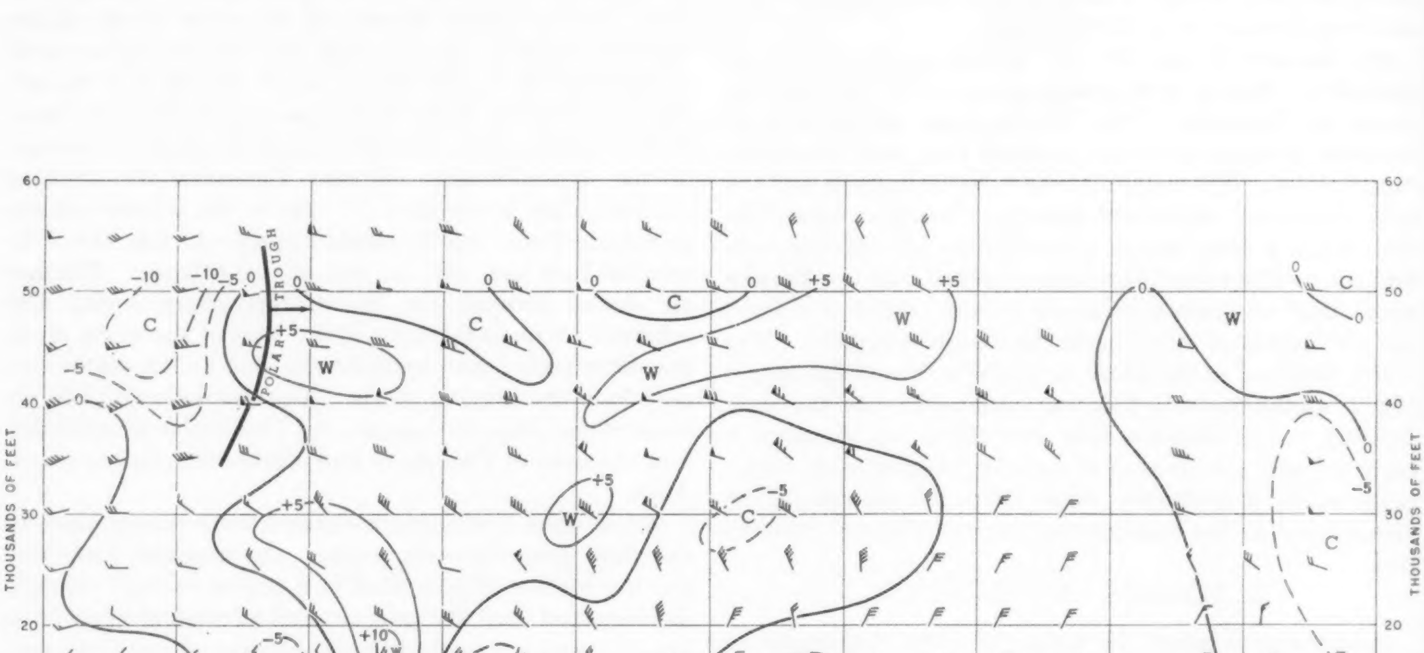

FIGURE 12.—Time section at San Juan, P. R., December 25, 1954–January 3, 1955.

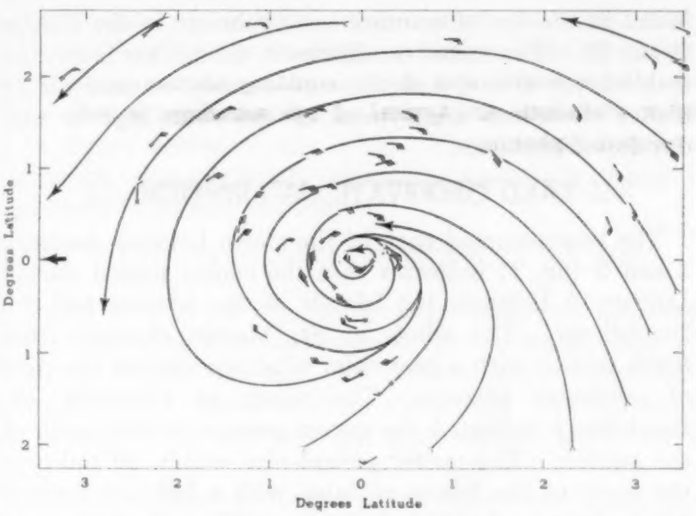

FIGURE 13.—Composite wind map and streamlines for January 2, 1955. Wind observations (in knots) between 0000 and 2100 GMT are plotted relative to storm position with direction of motion indicated by heavy arrow at left side of chart. Land observations marked by X; aircraft observations, by square; others are ship observations. Wind speed in knots. Units of distance in degrees of latitude.

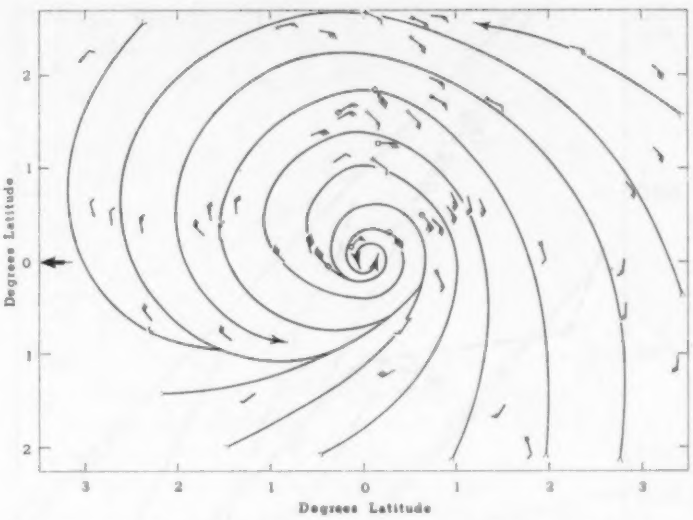

FIGURE 14.—Composite wind map and streamlines for January 3, 1955.

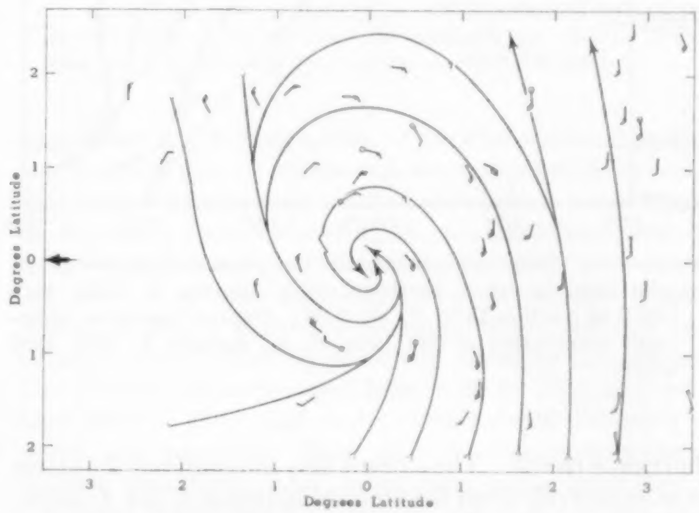

FIGURE 15.—Composite wind map and streamlines for January 4, 1955.

the source of polar air and instead a current from the SE from the tropical Atlantic entered the circulation of the storm. Considerable cloudiness and widespread shower activity prevailed over the area from the beginning thus giving another source of heat by the release of latent heat. Favorable thermal conditions were thus established for the transformation into a warm-core type of circulation.

An unsuccessful attempt was made to determine the influence of the high tropospheric flow (200-mb. level) in the formation of Alice. Unfortunately, the observations at San Juan, Bermuda, and Ship Echo alone did not suffice for a satisfactory analysis in the area of the storm. The time cross-sections of San Juan and Bermuda (figs. 11 and 12) give an indication of the flow prevalent at high levels during the period. The polar trough aloft passed Bermuda around 0000 GMT, December 26; the tropical extension passed San Juan around 1500 GMT, December 26. It moved eastward and passed Ship Echo around 1500 GMT, December 27. It was established in the region to the east of Ship Echo extending southwestward to the region of the Lesser Antilles around December 28 and 29 at the time that fracture and cyclogenesis occurred at lower levels. After the passage of the trough a strong west to west-northwest current prevailed over the tropical latitudes (see the San Juan upper winds). It was impossible to determine the presence of any jet structure in the flow, but the maximum wind speeds at San Juan were all the time at least 50 knots.

4. WIND AND THERMAL STRUCTURE

Observations obtained during January 1–4, 1955 indicate that Alice developed true hurricane properties. There are two essential characteristics that may be considered as typical of tropical hurricanes. One is the organization of the wind system with the maximum winds in a ring close to the center and speeds decreasing radially outward. The other is the warm-core type of thermal structure. Both of these properties were attained by Alice in its relatively short life.

For a better picture of the wind field, composite wind maps were prepared. These charts were obtained by plotting all available wind observations relative to the center of the storm for the period 0000 to 2100 GMT each day. Figures 13, 14, and 15 indicate the composite charts for January 2, 3, and 4. The charts for January 2 and 3 reveal the typical circulation of a tropical cyclone. The strong winds of 50–60 knots were all observed very close to the center. Winds of 30–40 knots were recorded within 75 miles from the center. The right semicircle apparently was the strongest in accordance with results found in tropical storms. The streamline field was also typical of

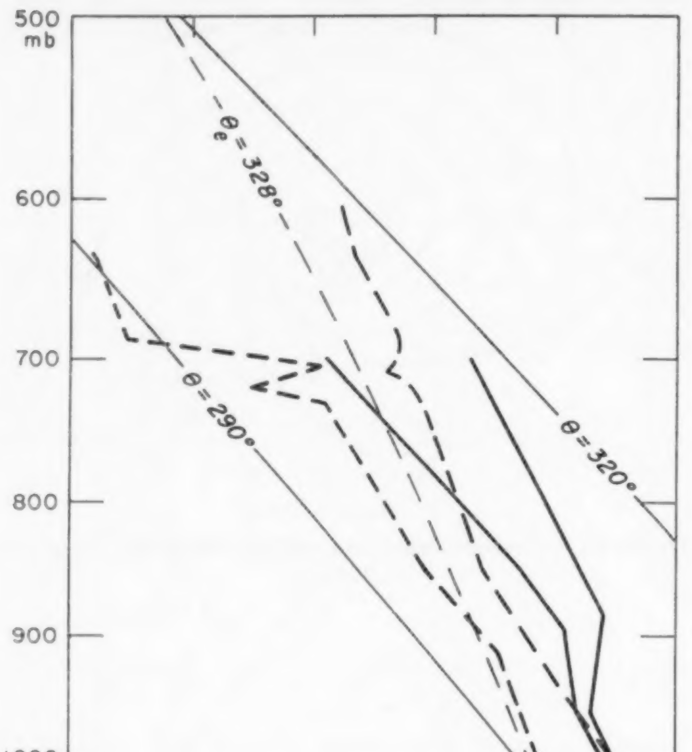

FIGURE 16.—Dropsonde observation (temperature and dew point, solid lines) in eye of hurricane Alice, January 3, 1955, 1345 GMT at position 16.5° N., 64.4° W. Dashed lines show radiosonde observation at San Juan, P. R., January 3, 1955, 1500 GMT.

hurricane charts. The current into the center of the storm was exclusively from the tropical latitudes of the Atlantic.

The chart for January 4 shows a different picture. During this day the storm was already in the dissipating stage. Wind speeds of 15–25 knots prevailed over most of the area. Only three winds of 30 knots were recorded: two from ships and one from the reconnaissance plane.

An Air Force reconnaissance report on the storm on January 3 described the eye as horseshoe shaped, poorly defined, 15 miles in diameter, with minimum pressure of 999 mb. A sounding in the eye obtained by this flight is plotted in figure 16 together with the observation that was made at San Juan about the same time and which was representative of the outside air. The warm-core properties of the storm are evident. The air of the eye was only slightly warmer than the outside air in the layer near the surface, but averaged about 6°–7° C. warmer from the 950-mb. level up. A nearly dry-adiabatic layer existed near the surface topped by a fairly deep stable layer. The lapse rate above the stable layer was slightly more unstable than the moist adiabatic. The eye conditions indicated by this sounding agree quite well with those found in the eye of summertime typhoons in the Pacific Ocean [2]. The mixed conditions in the surface layer, the stable layer above, and the considerably warmer air at higher elevations are typical of eye soundings in well developed typhoons.

5. LAND OBSERVATIONS AND DAMAGE

The reconstructed path of the storm between January 1 and 3 (fig. 1) indicates that the center passed during January 2 between the islands of St. Martin and St. Barthélemy. The winds at St. Martin changed from north to east with a near calm wind reported at the time of minimum pressure. The winds at Gustavia, St. Barthélemy indicated the center passage to the north of the station. The center passed also within 10 miles to the north of the Island of Saba, with a lull and gradual wind change observed between 1600 and 1900 GMT. Estimates of the maximum winds were not available at all places. At the Island of Saba an unofficial estimate of 75 m. p. h. was reported. A summary of the wind observations and damages is given in table 1. This is based on post-storm reports received from the governments of these islands. Most of the damage occurred within 50 nautical miles of the center. At St. Croix, which came within 65 nautical miles of the center, the effects of the storm were nil. The same occurred at the Island of Nevis, 50 nautical miles on the east side (left-hand semicircle of the storm).

The rainfall amounts were quite excessive in most places. The maximum report was made at Saba where orographic effects played a significant part. Readings of 6 to 8 inches were made at some of the other islands. The report of 2.48 inches at St. Martin looks surprisingly low, compared to the others. However, since the rainfall distribution in such storms is not uniform, the report is within the realm of possibility.

TABLE 1.—*Summary of wind, pressure, and rainfall observations and damage estimates in the Leeward Islands during hurricane Alice, January 1955*

Station	Average distance from center (miles)	Wind observations	Minimum pressure (mb.)	Rainfall (inches)	Estimated losses
St. Martin (Dutch section).	5	7½ hours of winds over 38 m. p. h.; no estimate of maximum.	991.2	2.48	
Gustavia, St. Barthélemy.	8–10	Maximum wind: south at 35 kt.	991.0		$43,000.00
Saba	6	Maximum wind: (estimated) 75 m. p. h.	982.5 at station, 1,500 ft. elev.	11.27 in 48 hr.	280,000.00
Anguilla	14	12 hours of winds over 38 m. p. h.		6.75 in 20 hr.	244,500.00
St. Eustatius	22		1001.4	8.00 in 40 hr.	31,000.00
St. Kitts	40	SW, force 6–8	1002.4	6.05 in 23 hr.	25,000.00
Nevis	52			3.60 in 19 hr.	Nil
Antigua	78	South, force 5	1007.1	2.00 in 8 hr.	Nil
Barbuda	52		1005.8	2.95	Nil
St. Croix	66				Nil
Total					623,500.00

Estimates of damage totalled over $600,000. The damage was mostly to shipping facilities and to crops. According to reports, the damage was caused mostly by the effect of rainfall and the action of the sea and not by the direct action of the wind. These total damages, although relatively small, represented a severe blow to the economy of these small islands.

6. OTHER COLD-SEASON TROPICAL STORMS

Alice of 1955 was the first tropical storm ever recorded officially during the cold season in the Caribbean-Atlantic area of tropical storms. As a result of its formation, several reports about previous wintertime tropical storms in history were revived. All such reports which have come to the attention of the writer referred to storms in the North Atlantic and most of them were very uncertain and indefinite. One particular report that received wide distribution in the newspapers concerned the storm experienced by Christopher Columbus during February 12–15, 1493. This storm, however, apparently was not a true tropical storm. At the time, Columbus was well advanced on the return leg of his first voyage to America. He sighted the Azores Islands on February 15, 1493; therefore, Columbus experienced the stormy weather in the region immediately to the west of the Azores. C. F. Brooks [3] made a detailed study of this storm and of another that affected Columbus on the same trip during February 26–March 4, 1493 and concluded that the weather and wind changes described by Columbus could have been produced by an extratropical storm with frontal boundaries such as occur frequently in that region today. Columbus had several experiences with hurricanes in his later voyages, but all were during the so-called regular hurricane season.

Although Alice, 1955, has been recorded as the first wintertime tropical storm in modern times, a closer study revealed that it was not the first one of its kind. In fact, one doesn't have to go very far in history to find a previous case of a storm of tropical characteristics during the cold season. Alice of 1955, for one thing, provided ample proof that they do occur.

During January 1951 a storm that occurred in the tropical regions of the Atlantic Ocean north of Puerto Rico gave all the indications of being as good an example of a tropical storm as Alice was. Figure 17, the surface chart for 1230 GMT, January 6, 1951, shows the storm centered at 21° N., 61.5° W., with a minimum pressure of around 1004 mb. and a wind report of force 8 (37 knots). Six hours after this map a ship reported winds of force 10 (48–55 knots) close to the center. This storm started as a frontal wave cyclone during January 1 around 34° N., 69° W. It moved first eastward and then curved southward toward tropical latitudes. By January 3 it was already isolated from the polar air and no fronts were distinguished in the circulation after this date. It continued in a southward track reaching latitude 20° N.

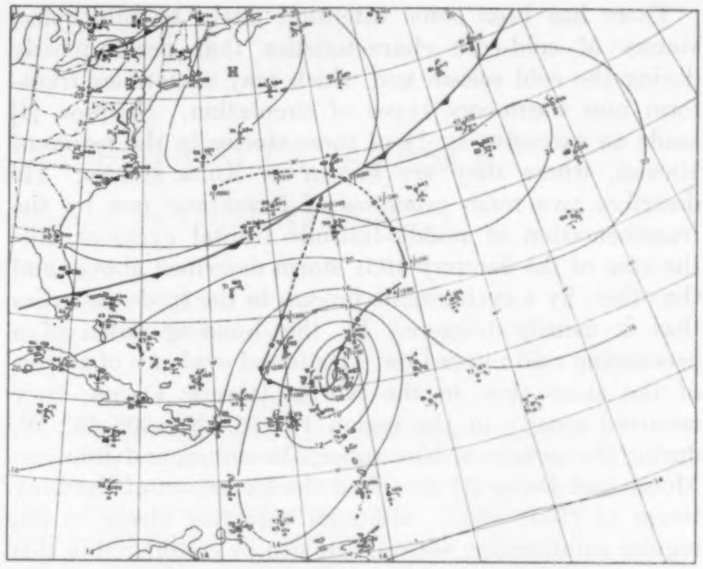

Figure 17.—Surface map, January 6, 1951, 1230 GMT. Track of the storm during the period that it exhibited frontal boundaries (extratropical characteristics) is indicated by dashed lines; track while it exhibited tropical characteristics is solid.

on January 6. A recurvature to the west occurred at this point, just in time to prevent the storm from striking land. This storm later turned northward, joined a polar front. on January 10 and became again an extratropical storm.

Between January 4 and January 9, 1951, this storm exhibited properties resembling a true tropical storm. Winds of force 7 to 10 (28–55 knots) were reported, the stronger winds generally observed closer to the center. The central pressure ranged from 1000 to 1005 mb., perhaps lower. The wind and pressure fields indicated a warm-core structure. Had aircraft observations been made in this storm, they undoubtedly would have shown a tropical warm-core circulation; and if the storm had struck land, it would have taken first honors as a cold-season hurricane instead of Alice of 1955.

A cursory investigation of the conditions for the formation of the storm of January 1–10, 1951 revealed some of the favorable circumstances present in the case of Alice 1955. Isolation of the Low from the belt of the westerlies both at the surface and at the 500-mb. level was also accomplished in the 1951 case. (See the *Daily Series Synoptic Weather Maps, Northern Hemisphere* for January 1951.) A blocking ridge to the northwest of the storm prevented subsequent intrusions of cold air while the storm wandered southward over warmer ocean waters and attained apparent warm-core characteristics. The original formation, however, was different from the 1955 case. In the January 1951 storm the initial Low was definitely formed as a frontal wave. Then the fronts vanished gradually and left a tropical type of Low. In the case of Alice, the initial formation occurred through a cyclogenetic process in the easterly current. The storm formed and moved all the time in tropical latitudes.

There has been some attention lately to subtropical storms of cold-core characteristics that occur mostly during the cold season and which may sometimes transform into warm-core types of circulation. Simpson [4] made an extensive study of these storms in the region of Hawaii, where they are known as Kona storms. He describes two main processes of formation: one by the transformation of middle-latitude frontal cyclones, like the case of the January 1951 storm described above; and the other, by a cyclogenetic process in the trade easterlies that is usually triggered by the building down of a preexisting cold upper Low. He found evidence of storms of the same type in the North Atlantic Ocean; they occurred mostly in the region 15°–35° N., 30°–60° W. during the months of November, December, and January. Moore and Davis [5] described the formation of the May storm of 1951, which, although occurring closer to the regular summertime season, can also be considered in this same class.

7. CONCLUSION

In summary, Alice of 1955 was a small storm. Its maximum winds barely reached hurricane intensity; its diameter reached no more than 50–60 nautical miles; it resembled to some extent some of the preseason and postseason storms, that is, the May and June storms and some of the October–November storms. The vertical extent was also very limited. It is perhaps significant that intensification occurred in a region which is also preferred for formation during the summer season. Alice represents a type of storm for which we have to watch more closely in coming years. Such storms are not as infrequent as we have been inclined to believe. It is important that forecasters recognize their true nature at the earliest time. The forecasting of their motion and intensity and issuance of adequate warnings to threatened areas would be a lot easier if we recognize them for what they really are.

ACKNOWLEDGMENT

The encouragement and assistance received from Mr. Ralph L. Higgs, United States Weather Bureau, San Juan, Puerto Rico, during the course of the study and preparation of this report is gratefully acknowledged.

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STRONG SURFACE WINDS AT BIG DELTA, ALASKA

An Example of Orographic Influence on Local Weather

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ABSTRACT

The remarkably high frequency of strong surface winds in the region of Big Delta, Alaska, is studied with respect to its cause, characteristics, and local effects. During the winter, the winds are predominantly east-southeast and, unlike glacier or valley winds, are caused by a topographically induced convergence of the flow of air down the Tanana Valley which occurs at times of southeast gradient winds aloft. Strong south winds are also experienced the year round. A noteworthy characteristic of the east-southeast winds is their persistence; an extreme case is described in which gusts in excess of 40 m. p. h. endured for 7½ days (January 20–28, 1952). Another characteristic of these winds is the marked diurnal variation in the frequency of their commencement, by which a strong control by atmospheric tides is inferred. An important effect of the winds is to interrupt periods of very low temperature, but sometimes to create severe "wind chill." The paper concludes with a brief account of the forecast problem.

1. INTRODUCTION

Big Delta, located about 85 miles southeast of Fairbanks, Alaska, experiences a remarkably high frequency of strong surface winds. The weather records maintained there in recent years by the Civil Aeronautics Administration afford some insight into the cause, the characteristics, and some of the local effects of these winds. A study of the Big Delta winds is described in this paper which was prepared by the author preliminary to the development of an objective forecast study of the winds. Practical interest in the region is chiefly due to the presence of the United States Army Arctic Test Branch, which is charged with the testing of military supplies and equipment for adequacy and satisfactory performance under conditions of extreme Arctic cold; the winds comprise one of the more troublesome forecast problems in Alaska inherited by the Air Weather Service.

Surface wind roses for each month of the year at Big Delta are presented in figure 1. These serve to emphasize the following: (1) A very high frequency of east-southeast winds occurs in the colder part of the year (October through March), these winds being relatively strong. (2) An important frequency of *southerly* winds prevails throughout the year. (3) The percentage of calms during the winter months averages about 13 which may be contrasted with an average of more than 50 percent at nearby locations such as Fairbanks and Eielson Air Force Base which have nearly identical climatic temperature regimes.

[1] Paper presented at the 6th Alaskan Science Conference, College, Alaska, June 1955.
[2] Based on a study by the author when on active duty with the Air Weather Service. Verbally cleared for publication by Information Services, AWS.

2. CAUSE OF WINDS

The cause of the anomalously high frequency and speed of the east-southeast and south winds at Big Delta can readily be traced to the characteristics of the terrain there. Figure 2 delineates the general topography over southern Alaska and extreme western Canada. While Fairbanks is located in the broad Tanana Plain where the terrain is rather level, Big Delta is located at the southeast edge of the Plain, near the mouth of the long, relatively narrow valley in which the Tanana River flows from its headwaters. This valley is oriented approximately west-northwest—east-southeast, and while its floor rises gently about 1,000 feet from Big Delta to Northway, it is paralleled on either side by mountains which rise 2,000 to 7,000 feet higher.

When the weather pattern is such as to induce a southeasterly flow over the valley, and to set up a surface pressure gradient between Northway and Big Delta (with the higher pressures at Northway), air is funneled down this valley toward Big Delta with convergence sufficient to result in a surface wind current of strongly supergradient speed. This current, which has been likened to a jet stream against the ground, has been described by Ehrlich [1] on the basis of special aircraft reconnaissance in 1951. When blowing snow carried along by the wind makes the "jet" visible, it is apparent that (*a*) it is about 5 miles wide near Big Delta and perhaps 2,000 feet deep against the ground, and that (*b*) it does not break up as it enters the broad Tanana Plain northwest of Big Delta, but often continues across the plain like "a stream of water coming from the nozzle of a hose," and may stretch for more than 100 miles as it gradually broadens and diminishes in speed.

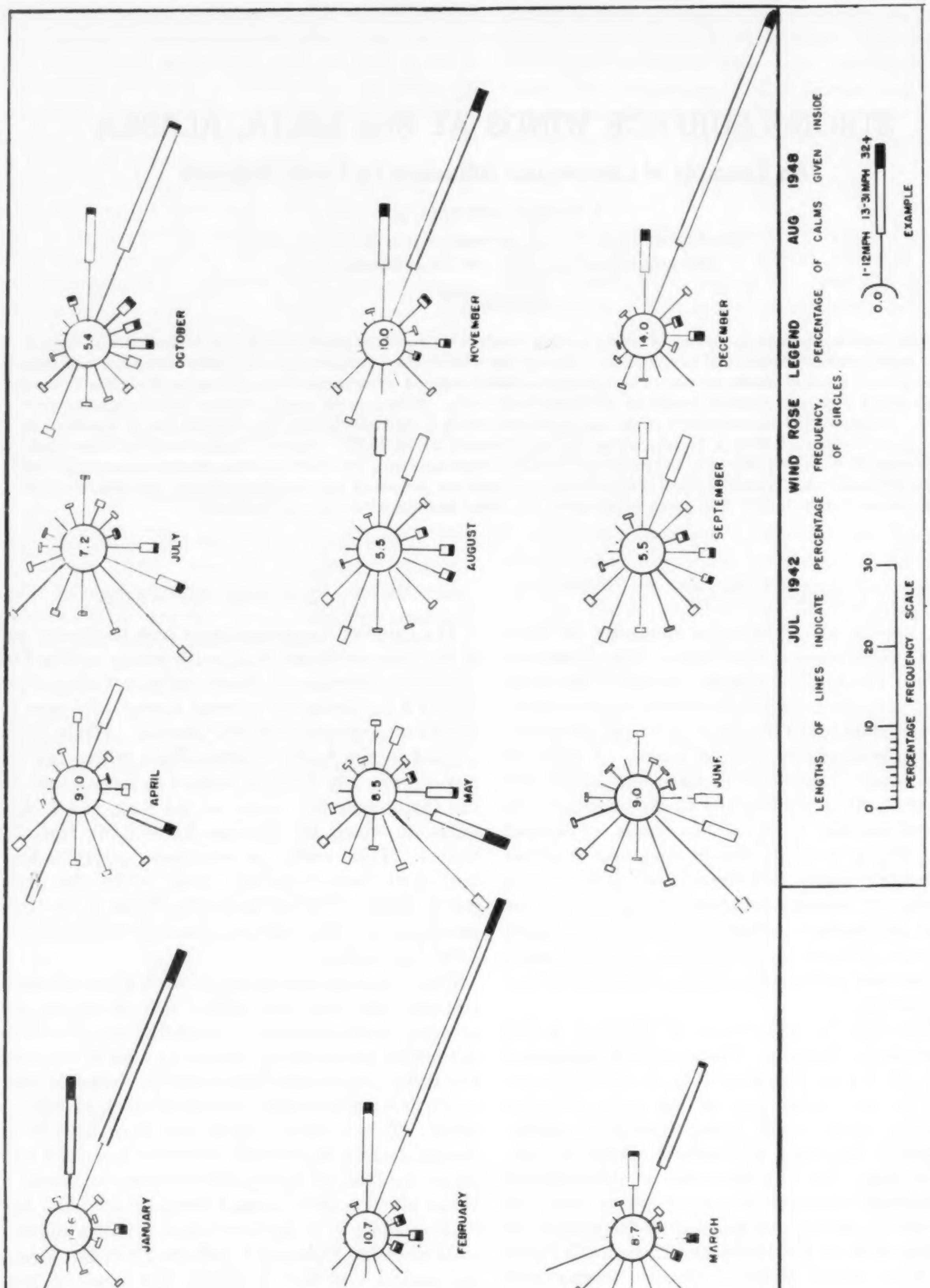

FIGURE 1.—Monthly wind roses for Big Delta, Alaska, based upon records from July 1942 through August 1948. Duplicated from a portion of a published climatology chart of the 7th Weather Group, Air Weather Service.

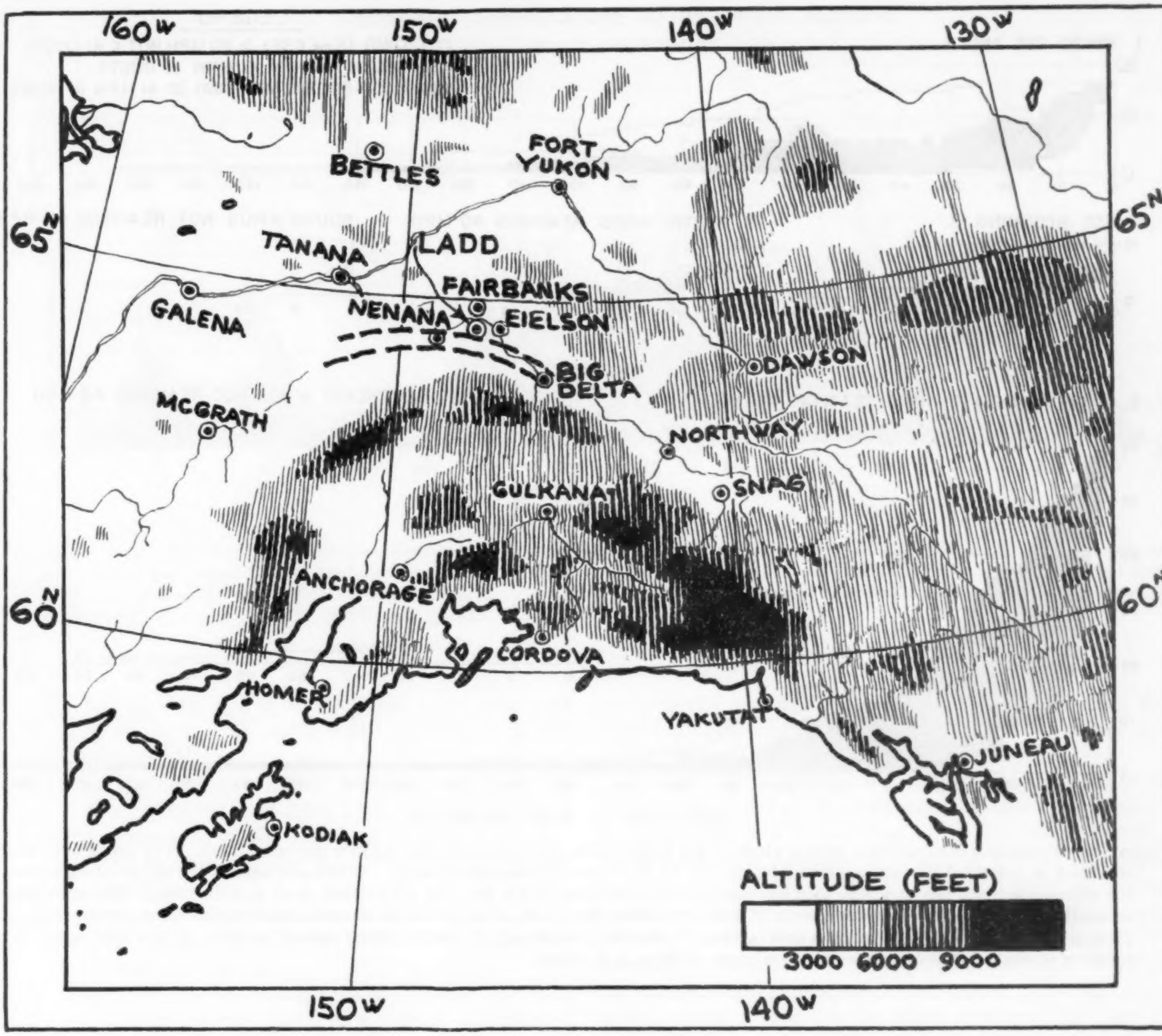

Figure 2.—Topography of south central Alaska. Dashed curves delineate the jet-stream-like surface wind in the position observed by aircraft reconnaissance in 1951 (Ehrlich [1]).

The "jet" has sometimes been observed to pass over Nenana, 110 miles to the west-northwest, where it has caused strong, gusty surface winds. On such occasions, Ladd and Eielson Air Force Bases located a few miles to the north of Nenana have experienced no wind whatever (note the "jet" position in fig. 2).

It is known that this current of air does not pass directly over Big Delta and the adjacent Arctic Test area with every occurrence. Slight differences in the pressure field south and east of the area cause the wind stream to vary its course and sometimes to bypass it. Moreover, it is likely that the "jet" meanders, and on that account gives rise to large variations in the strength of the surface wind at Big Delta.

To the south of Big Delta, at a distance of about 20 miles, lies a formidable portion of the Alaska Range which is locally interrupted only by a narrow pass, oriented north-south, through which flows the Delta River. In the pass, the adjacent mountains rise between 6,000 and 10,000 feet above the river; when the general weather regime results in a southerly or south-southwesterly flow over the Gulkana Basin to the south, a stream of air is usually forced through the pass, attaining highly super-gradient speed. This southerly current rather seldom passes directly over Big Delta. Owing both to the usual development and movement of pressure systems in Alaska, and to the exact orientation of the mountain pass itself, the current generally moves out into the Tanana Plain at

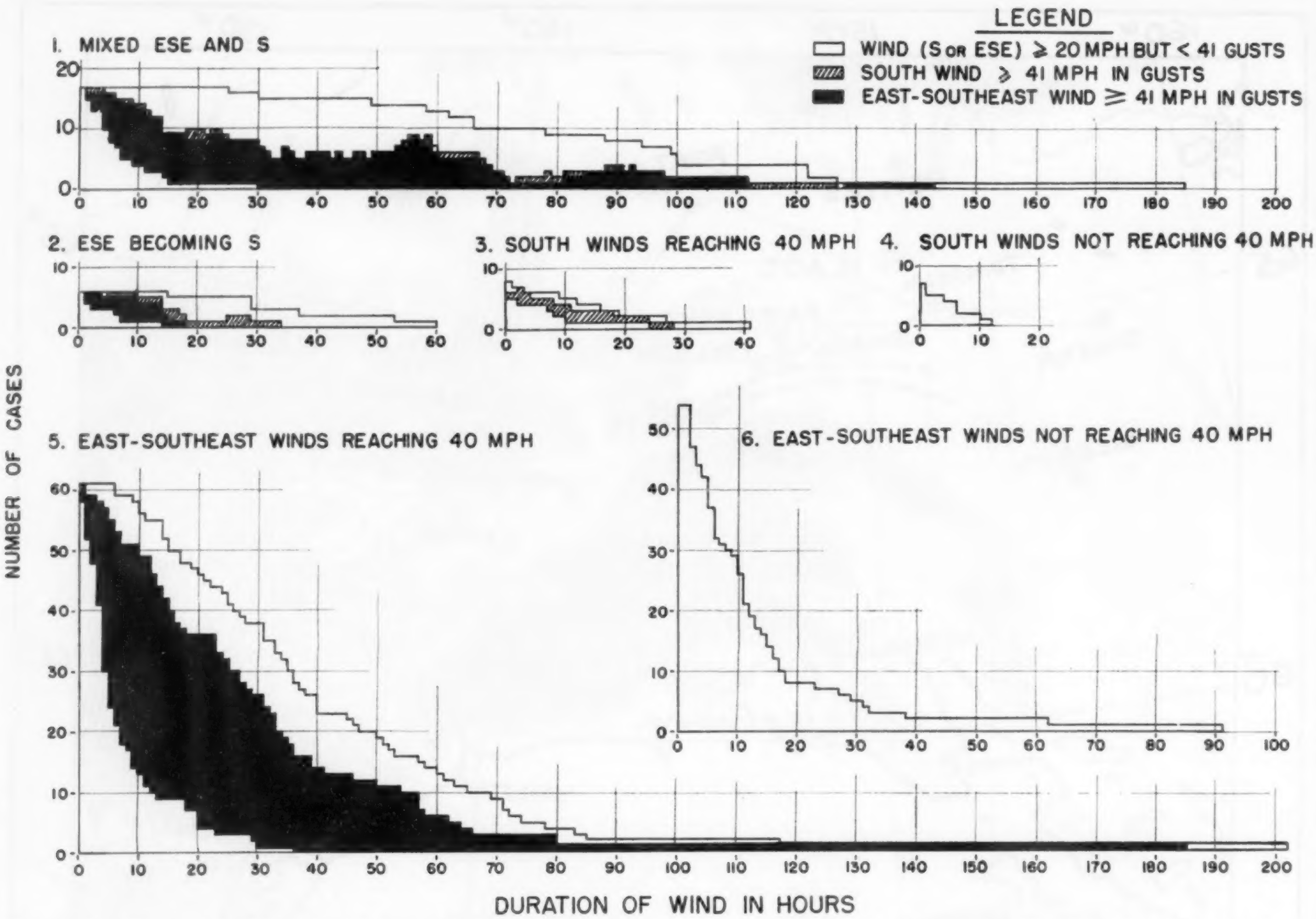

FIGURE 3.—Histograms of duration of strong winds at Big Delta during the colder months October through March. The winds have been classified, as indicated, into six categories determined by their direction and speed range. White area against the left-hand edge shows the decrease in frequency of winds, with increasing time of duration, which have not yet reached 40 m. p. h. in gusts. The solid black area adjoining shows the relative frequency of winds exceeding 40 m. p. h., as a function of elapsed time since the commencement of the 20-m. p. h. winds, The trailing white area reflects the frequency of 20-m. p. h. winds which followed the 40-m. p. h. winds, again as a function of elapsed time since the first occurrence of 20-m. ρ. h. winds.

a point lying to the west of Big Delta instead. The high incidence of strong *south* winds in the area must evidently be explained otherwise: Evans [2] has concluded that a strong, föhn-type flow of air directly over the Alaska Range is probably involved along with, or instead of, the current of air passing through the Delta River Valley. In support of this contention, Evans has observed that strong south winds at Big Delta are accompanied by a föhn-wall cloud formation over the mountains. Also, to be discussed presently, such south winds are accompanied by a very great temperature increase at the surface.

The weather situations giving rise to the strong east-southeast winds are not very different than those responsible for the strong south winds. When east-southeast winds are prevailing at Big Belta, south winds are often occurring just to the west, and a zone of convergence between the two currents can be found in the vicinity. These circumstances have been witnessed on several occasions by Evans as he was travelling over the network of roads in the Arctic Test Branch area. The south current is sometimes observed to encroach steadily into the region previously occupied by the east-southeast current, with the result that the reported wind at Big Delta may shift abruptly as the result of a gradual trend in the regional weather situation.

3. CHARACTERISTICS OF WINDS

A characteristic of the Big Delta winds which is worthy of particular attention is their duration. For purposes of describing duration, it is advantageous to separate occurrences of strong winds (defined as greater than or equal to 20 m. p. h.) into six categories:

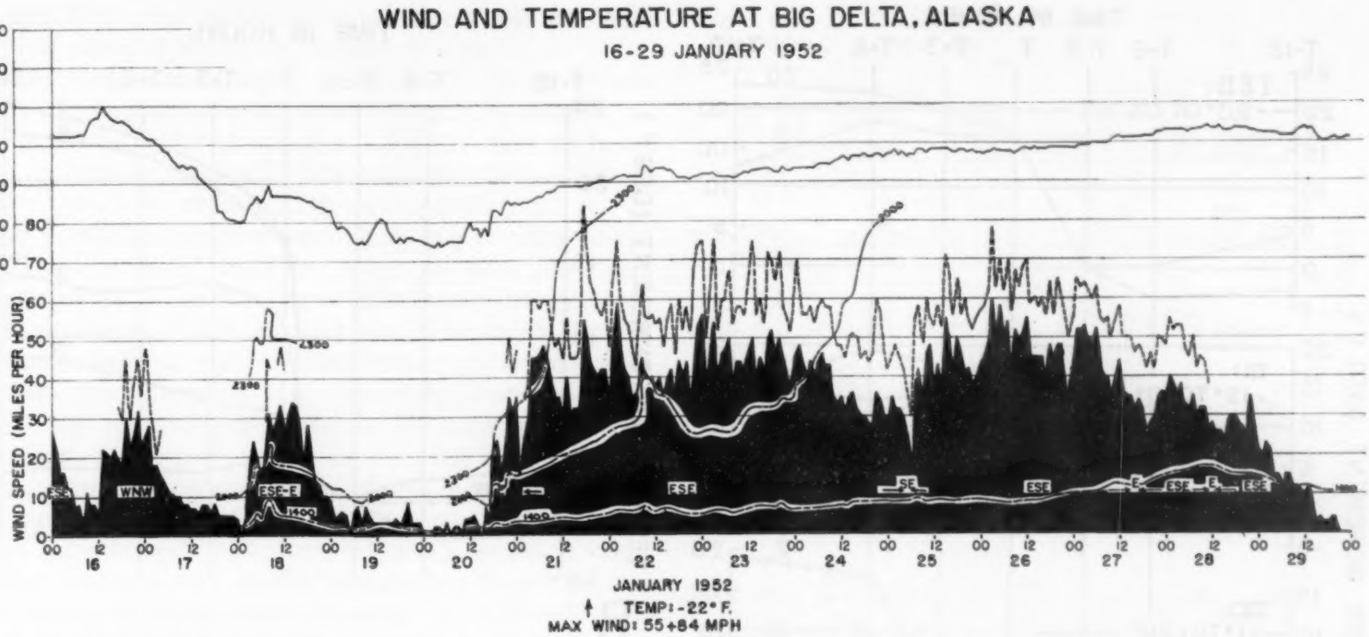

Figure 4.—Example of protracted duration of strong east-southeast winds at Big Delta showing variation in speed (fastest minute and gusts, which are dashed peaks), wind direction, temperature, and wind-chill factor. The wind-chill factor (sloping lines labeled 1400, 2000, and 2300) is plotted on the wind-speed graph as a function of concurrent temperature in order that the influence of the wind on human comfort during the storm can be readily seen.

(1) Mixed ESE and S winds (with gusts exceeding 40 m. p. h. from one or both directions, and usually both).
(2) ESE winds becoming S (usually with gusts exceeding 40 m. p. h. from both directions).
(3) S winds only which exceed 40 m. p. h.
(4) S winds only, greater than 20 m. p. h. but not exceeding 40 m. p. h.
(5) ESE winds only, which exceed 40 m. p. h.
(6) ESE winds only, greater than 20 m. p. h. but not exceeding 40 m. p. h.

In figure 3 are shown histograms of duration for each of the 6 individual classes, based on a total of 153 cases of wind in the four 6-month winter seasons 1949–50 through 1952–53. In these diagrams, the commencement of a wind is defined as the time of first observation (at the CAA station) of wind speed equal to or greater than 20 m. p. h. A few cases in which the wind reached 20 m. p. h. in a single, isolated observation have not been included in the data. Figure 3 serves several purposes. First, it shows that the most common type of wind (according to this classification) is that which blows invariably from the east-southeast, and which at some point in its history is likely to exceed 40 m. p. h. Second, it reveals the duration spectrum for each of the various categories of wind; e. g., in the case of east-southeast wind reaching 40 m. p. h., about 20 percent are still blowing above 40 m. p. h. at the end of 48 hours following the commencement of the wind. Third, figure 3 gives the probability that an existing wind will terminate or change in speed category in a given period of time, which can be applied both in the planning of operations affected by the strong winds and in the forecasting of temperature at Big Delta.

Occasionally an east-southeast wind of exceptional duration occurs. The most notable case of this type to occur in recent years was that of January 20–28, 1952, illustrated in figure 4. On January 20, a strong surface anticylone located over the Mackenzie Basin was further intensified by the building up of a blocking ridge over the Bering Strait; the wind at Big Delta commenced from the east at 1800 Alaska Standard Time (0400 GMT, January 21). By 2100 AST the same day, gusts had reached 44 m. p. h. Then the surface wind veered to the more common direction of east-southeast, and with the exception of two periods of only two hours each, east-southeast winds gusting in excess of 40 m. p. h. persisted until 1000 AST on January 28. As is typical with the onset of east-southeast winds (see below), the temperature rose on January 20 by 22 Fahrenheit degrees from a low of $-52°$ F. During the following 8 days, the temperature continued to rise rather slowly to a high of $+8°$ F. on the 28th. The combination of high winds and relatively low temperatures made life at Big Delta rather hazardous. For the first 15 hours of wind, the wind-chill index [3, 4], which reflects the loss of heat from human skin in terms of cal/m^2 hr., was more than 2300. An index of 2300 is the point at which exposed areas of the face of an average person will freeze within 30 seconds. And for the first 86 hours (3½ days), the wind-chill index was continuously more than 2000 (the value at which a human face requires one minute to freeze,

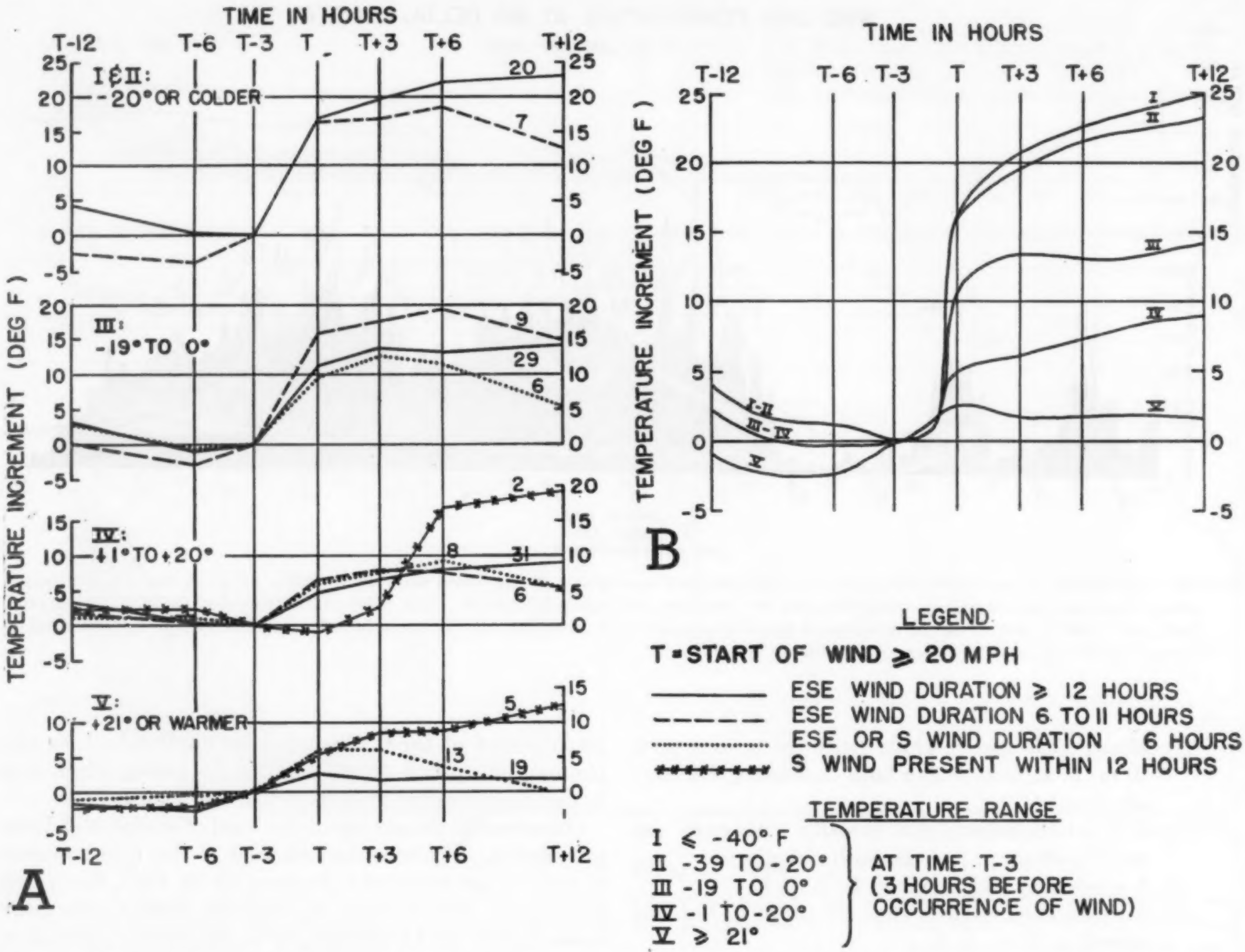

FIGURE 5.—Effect of winds on temperature at Big Delta, Alaska. The left side (A) shows, for each of five ranges of initial (3 hours before onset of wind) surface temperature, the temperature change caused by the inception of east-southeast and south winds of various durations as indicated. The right side of the figure shows, for the same ranges of initial temperature, the temperature change caused by the inception only of east-southeast winds which endure 12 hours or more. Here, the temperature changes have been interpolated more realistically between the computed data.

and travel or life in temporary shelter is described as dangerous). And for a period of 9 days, which included the period of strong winds, danger of frostbite was continuously present with the wind-chill index remaining above 1400.

4. LOCAL EFFECT OF WINDS ON TEMPERATURE

As mentioned previously, an important consequence of strong winds during the winter at Big Delta is normally to interrupt periods of low temperature. The wind is such a common occurrence there that the inhabitants take it rather in their stride. Too, the Army Arctic Test Branch is involved more with "cold-soaking" tests than it is with wind-chill or high wind, and when a wind impends, tests are usually suspended merely because the temperature will be increased above the desired level.

Figure 5 illustrates the relationship between wind and temperature. The first increase of wind to 15 or 20 m. p. h. causes an abrupt increase in the surface air temperature, presumably through the partial destruction of the Arctic inversion in the region by vertical mixing. Further increases of wind above 20 m. p. h. have the effect of very slight additional warming when the wind is east-southeast, but of strong warming when the wind is southerly.

The left-hand side of figure 5 shows the increase in temperature with the onset of winds equal to or greater than 20 m. p. h., for each of several ranges of prewind surface temperature. The different lines in the graphs identify the effect of four categories of wind defined as follows:

(1) ESE wind enduring at least 12 hours (whether or not they exceed 40 m. p. h.).

(2) ESE wind enduring for between 6 and 11 hours (some of which may exceed 40 m. p. h.).
(3) ESE wind enduring for less than 6 hours (sometimes with a brief period of southerly wind).
(4) Wind which becomes southerly within 12 hours (which starts as ESE, but which usually increases to above 40 m. p. h. from the south).

For particular temperature ranges, one or more of the above categories of wind are found to exist too infrequently to yield a statistically reliable average temperature trend, and these have been omitted. This part of figure 5 demonstrates that, with low temperatures before the onset of wind, there results a larger warming than with relatively high prewind temperatures. The duration of the wind plays a secondary role in determining the magnitude of the warming, and appears to affect the temperature to varying degrees chiefly after the time when some of the winds have ceased. The case of south winds is noteworthy, inasmuch as a greater warming results with them. Most of the southerly winds occurred with relatively high prewind temperatures, but one case also occurred in each of the colder categories (II and III). These have not been included in the figure. It is interesting that, with the south wind which occurred in category II (prewind temperature −22° F.), a warming of 36° F. resulted in 6 hours, 46° in 9 hours, and 47° in 15 hours. The case of south wind in category III also resulted in a sharp warming trend; it is likely that the magnitude of the warming with southerly winds as a function of prewind temperature is more or less proportional to the magnitude of the warming with east-southeast winds, and larger.

The right-hand side of figure 5 refers specifically to east-southeast winds having a duration of 12 hours or more, which is the most frequent category encountered. In this part of the figure, the warming which results from the onset of wind is shown for various ranges of prewind temperature, facilitating a comparison of magnitudes. Here, the temperature trends have been interpolated between computed values in a manner which shows their shape more realistically. Three comments on this diagram are appropriate:

(1) Comparing the relative magnitudes of the warming trends in the separate temperature categories, it appears likely that the curve for category I represents an approximate upper limit to this magnitude. The mutual closeness of the curves for categories I and II cannot be explained in terms of a clustering of temperatures in category I about its upper limit of −40° F. (The three cases included in this category yielded an average prewind temperature of −47° F., and the 17 cases in category II, −30° F.).

(2) The dip in temperature for categories I and II just before the commencement of the wind can be explained as the result of excessive radiational cooling. It usually happens that a period of calm precedes the wind and endures for several hours.

(3) The gradual falling off of temperature following several hours after the onset of wind in the high-temperature category V is apparently because most of the cases in the category came in the late fall and early spring when there is a noticeable preference for the winds to commence in the late morning hours. Accordingly, a diurnal temperature trend has been introduced into the data.

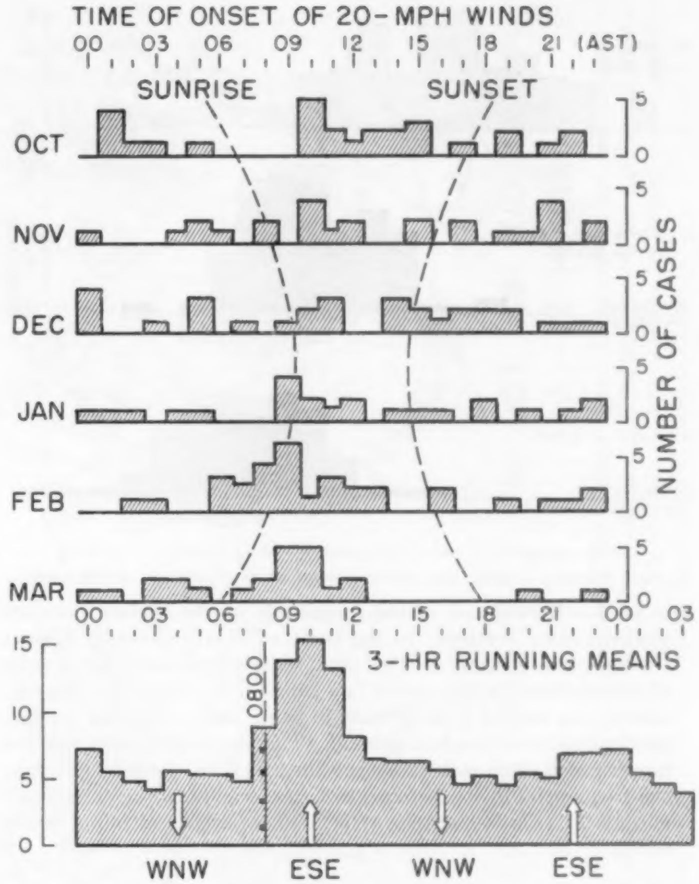

FIGURE 6.—Diurnal distribution of the time of onset of east-southeast winds at Big Delta (20 m. p. h. or more), in Alaska Standard Time. The upper part of the figure shows frequency distribution by months (4 years of data). The lower part shows the seasonal frequency distribution expressed as 3-hour running means. The arrows at the bottom of the figure denote the theoretically computed times of tidal accelerations which should affect wind frequency in the directions indicated.

5. DIURNAL EFFECT IN WIND FREQUENCY

Figure 6 shows, for all four 6-month seasons of data, the diurnal distribution of the first occurrence of wind from the east-southeast. It will be noted that in each month, with the possible exception of December, a preference for about 1000 AST exists. This tendency is brought out more forcefully by the seasonal totals of frequency as a function of the hour, which are shown at the bottom of the figure as 3-hour running means. The following remarks are pertinent to explaining this diurnal favoritism.

The approximate times of sunrise and sunset are indicated in figure 6 as a function of season. In contrast to the widely varying time of sunrise, the time of commencement of the winds is seen to be more or less constant

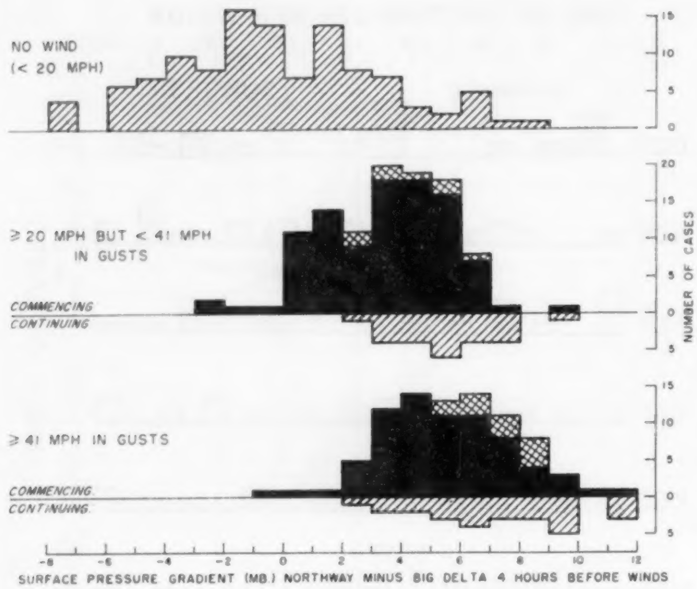

FIGURE 7.—Histograms of the magnitude of the surface pressure gradient from Northway to Big Delta which antecedes by 4 hours the inception of wind at Big Delta, with comparative histograms of the gradient for the case of "no wind" following (top of figure), and for the case of wind already in being and continuing (underportions of lower two histograms). To obtain the correct relative number of cases in each category shown, it is necessary to multiply the frequencies in the hatched histograms by a factor of approximately 20. Cross-hatching refers to frequencies of initial south winds.

through the season. Moreover the sun is at such a large zenith angle during the middle of the winter that very little heating within the Arctic inversion can take place until the late morning.

At about 0800 local time, the sun is nevertheless able to reach the floor of the Tanana Valley between Northway and Big Delta during every month. Even at the winter solstice, some heating within the Arctic inversion in the valley may be expected at that time, causing minimum stability within the layer shortly thereafter. Allowing for a further delay before the winds at the surface are increased by vertical flux of momentum and conveyed down the valley to Big Delta, one might explain the 1000 AST peak in the onset of winds there.

The secondary maximum in frequency of onset of the winds to be found 12 hours later (from 2200 to 2400 AST) suggests that a tidal influence is involved. Through his computation of the tidal wind fields in the lower atmosphere, Stolov [5] shows that, for the latitude of Big Delta, a tidal wind component of more than one m. p. h. from the east-southeast may be expected at about 1000 and 2200 local time. Since the tidal vector rotates clockwise with a period of 12 hours, a wind component from the opposite direction is expected at about 0400 and 1600 local time. The phase of this tidal acceleration is shown by the arrows at the bottom of figure 6 where it may readily be compared with the diurnal variation in the commencement frequency of the wind. A relationship is strongly suggested.

Accordingly, it is the writer's belief that the peak at 1000 local time in the frequency of onset of east-southeast winds is the result of a combination of a tidal component to the wind down the Tanana Valley which is locally magnified by topographic convergence, with a maximum of turbulent transfer of momentum in the Arctic inversion in the valley induced by solar heating of the layer shortly before that time.

6. FORECASTING THE WINDS

For a number of years, an index of the imminent likelihood of winds (especially the east-southeast type) has locally been employed in forecast practice, consisting of the sea level pressure difference between Northway and Big Delta. When the Northway pressure exceeds the Big Delta pressure by 3 mb. or more, winds exceeding 20 m. p. h. may be forecast to commence or to continue, as appropriate, with fair confidence. Figure 7 illustrates the degree of relationship between the Northway-Big Delta pressure gradient and the subsequent occurrence of wind at Big Delta, wherein the gradient anticipates the wind by 4 hours. It is well known that the winds are strongly cross-gradient, as these pressure differences imply them to be. They are much more strongly cross-gradient in the case of a cyclonic pressure regime over the area than they are in the case of an anticyclonic regime.

Figure 8 depicts the synoptic weather types which are associated with the onset of Big Delta winds. Types A through E have to do with east-southeast winds, and Type S involves south winds. For illustration, the actual surface and 500-mb. patterns for selected dates (extracted from the U. S. Weather Bureau Historical Map Series) are shown superposed. It will be seen that the range of synoptic patterns conducive to strong winds is rather broad. In figure 8, their classification has been set up on the basis of a common trend in synoptic development, although it is desirable to emphasize that each type is generally independent of the others.

Type A (top, left) illustrates a rather common type which gives rise to protracted periods of strong east-southeast winds. A blocking ridge aloft over western Alaska is associated with a strongly developed surface High over the Mackenzie Basin, which remains stationary. A strong gradient of pressure exists to the southwest of the Mackenzie High, and near-gradient flow exists over Tanana Valley above Big Delta. The extreme stability of location of the blocking pattern often results in long duration of the wind.

Type B (top, center) illustrates the initial breakdown of a blocking ridge, or alternatively the incipient formation of a block. In this case, a weak frontal cyclone moves from the west into the Gulf of Alaska to reinforce the Gulf Low. Local displacement of the Gulf Low toward a moderately strong Mackenzie High—sometimes in combination with the passage of the upper portion of the migrating front—results in a brief period of wind.

Type C (top, right) illustrates the case of a closed High cell over the Arctic Ocean which is connected by essen-

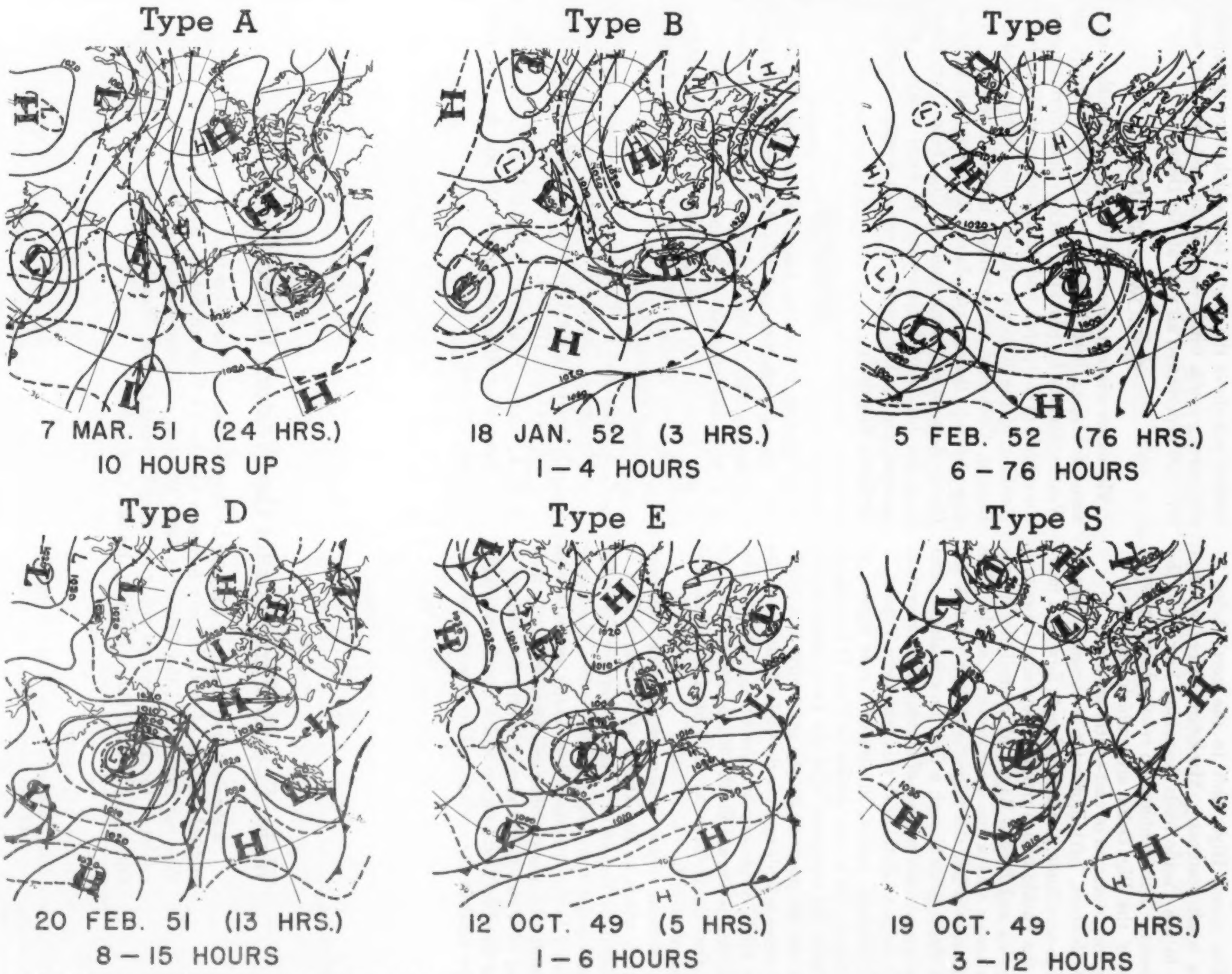

FIGURE 8.—Synoptic weather patterns in the vicinity of Alaska associated with the onset of strong winds at Big Delta. Types A through E are concomitant to east-southeast winds; Type S, to south winds. With each weather type is shown the date of the synoptic pattern used for illustration, the duration of wind at Big Delta on that occasion, and the usual range of duration of wind (> 40 m. p. h.) for that type. The 500-mb. pattern is shown by dashed contours at 400-foot intervals, and closed Highs and Lows aloft by small H's and L's respectively. Surface isobaric interval (solid lines) is 10 mb.

tially non-blocking ridges over the Rockies and/or eastern Siberia. The closed High aloft to the north helps to maintain a weak stationary Mackenzie High, while a zonal flow in the Pacific feeds deepening Lows into the Gulf of Alaska. The combination results in a tightened gradient over the Big Delta area which can result in moderately long duration of east-southeast wind.

Type D (bottom, left) illustrates a common case in which a large-amplitude, short-wavelength flow aloft moving across the Pacific brings rapidly deepening Lows into the Bering Sea and Bristol Bay, and intensifying migratory Highs across Alaska into northwestern Canada. The wind commences when the migratory High reaches the Mackenzie Basin, even though a Gulf Low is absent, and endures for a moderate number of hours. The movement of the High helps to distinguish this type from Type A.

Type E (bottom, center) illustrates a case typical of the late autumn. It is very similar to Type D in synoptic evolution with the exception that the Lows reaching the Bering Sea are not prone to deepen markedly, and consequently no ridge is built up over the Gulf of Alaska. The Mackenzie High, migratory or otherwise, is absent, and frontal systems penetrate northward across Alaska. In this type, the Big Delta wind is of brief duration, and occurs in advance of frontal passage.

Type S (bottom, right) illustrates the usual concomitant to south winds at Big Delta. Close parallelism to Type E is evident, but an important difference lies in the fact that the major storm center moving into the Bering Sea is favored to move more-or-less bodily across Alaska in a flow aloft which is stronger and more southerly than in the case of Type E. A strong southerly gradient is set up over the Big Delta region which usually results in a föhn-type wind at the surface. This synoptic pattern, in a somewhat weakened form, is very common during the warmer months of the year, a fact which serves to explain why strong south winds—unlike the east-southeast variety—are experienced all year at Big Delta.

ACKNOWLEDGMENTS

The writer wishes to record with appreciation the assistance of Maj. Albert Ehrlich, 7th Weather Group, Air Weather Service, for providing background information on the subject of this paper, and the assistance of Miss Alveda Nordling of the Regional Office, United States Weather Bureau, Anchorage, for providing microfilm records. He also wishes to thank Capt. James R. Evans, 7th Weather Group detachment commander at Big Delta (Fort Greeley), for his helpful discussions of local weather phenomena and for the supply of certain weather records to facilitate this study.

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THE WEATHER AND CIRCULATION OF JANUARY 1956 [1]

A Month With a Record Low Index

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1. FIVE-DAY MEAN CIRCULATION

In the preceding three articles of this series [1, 2, 3] the unusual behavior of the zonal index during the past three months has been noted. Five-day mean values of this index, which expresses the average strength of the prevailing westerlies between latitudes 35° and 55° N. in the Western Hemisphere, are plotted in figure 1 for both 700 mb. (above) and sea level (below). After undergoing an unprecedented double cycle in October 1955 [1], the index failed to rise above normal during any 5-day period from October 28 to February 22, the longest period of continuous low index ever observed. However, normal values were reached briefly, at 700 mb. for the period ending December 25 and at sea level for the period ending February 5. These two dates mark the beginning and end of a pronounced index cycle which culminated with the lowest 5-day mean indices ever observed, from January 7–11 at 700 mb. and from January 11–15 at sea level (table 1). In order to extend the period of record back beyond January 1941, when computation of indices for the Western Hemisphere began, the 5-day mean indices for these two periods were recomputed for the entire Northern Hemisphere. They were then compared with the hemispheric indices computed by Willett [4] for each 5-day period during the six cold months of

[1] See Charts I–XV following p. 45 for analyzed climatological data for the month.

TABLE 1.—*Values of the temperate latitude (35°–55° N.) zonal index (m. p. s.)*

	Period	Western Hemisphere only		Entire Northern Hemisphere	
		700 mb.	Sea level	700 mb.	Sea level
5-day mean	Jan. 7–11, 1956	[a] 2.9	−3.0	[b] 3.9	[e] −2.2
5-day mean	Jan. 11–15, 1956	3.4	[a] −3.4	5.5	−2.1
30-day mean	Jan. 1956	6.6	[d] 0.0	[e] 6.9	−0.5
Normal	January	11.8	4.1	9.6	2.0
30-day anomaly	Jan. 1956	[f] −5.2	[f] −4.1	[e] −2.7	[g] −2.5

[a] Lowest index for any 5-day mean period of record (1/41–1/56).
[b] Lowest index for any 5-day mean period of record (10/32–3/39).
[c] Third lowest index for any 5-day mean period of record (10/32–3/39 and 1/41–1/42).
[d] Lowest index for any month of record (9/42–1/56).
[e] Comparable data not available.
[f] Greatest negative departure from normal for any month of record (9/42–1/56).
[g] Greatest negative departure from normal for any month of record (1/99–6/39 and 1/49–1/56).

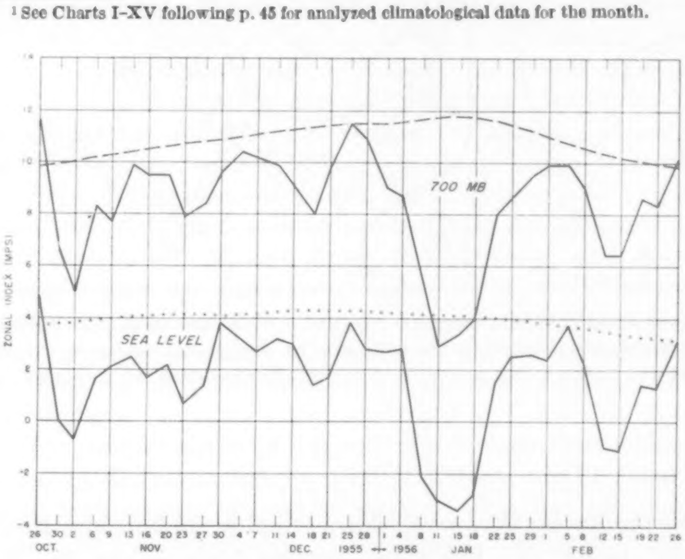

FIGURE 1.—Time variation of zonal index in meters per second for the Western Hemisphere in the latitude belt 35°–55° N. Solid lines connect 5-day mean values (plotted on the last day of the period) for 700 mb. (above) and sea level (below). Variation of normal zonal index is shown by dashed line for 700 mb. and dotted line for sea level. Note the prolonged period of low index (Oct. 28 to Feb. 22) and the marked index cycle from Dec. 25 to Feb 5.

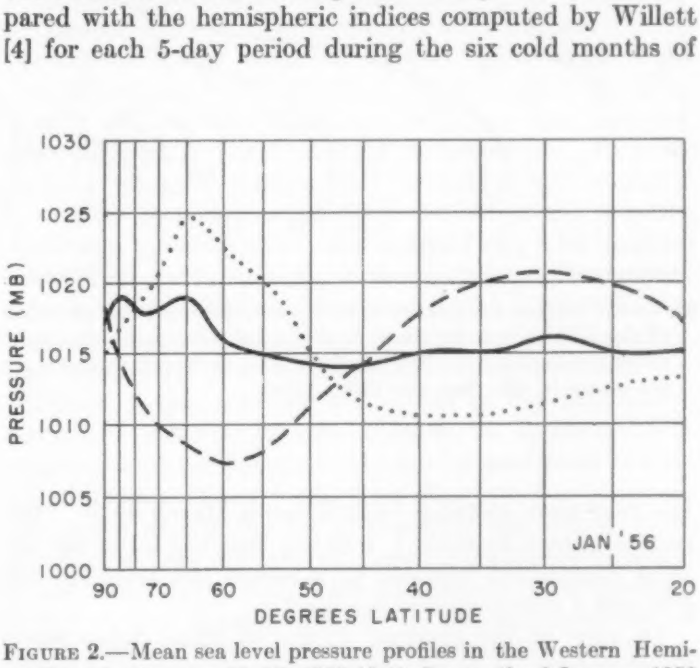

FIGURE 2.—Mean sea level pressure profiles in the Western Hemisphere for January 11–15, 1956 (dotted), month of January 1956 (solid) and January normal (dashed). Excess of pressure at high latitudes and deficit in the south were indicative of low zonal index during both 5- and 30-day periods.

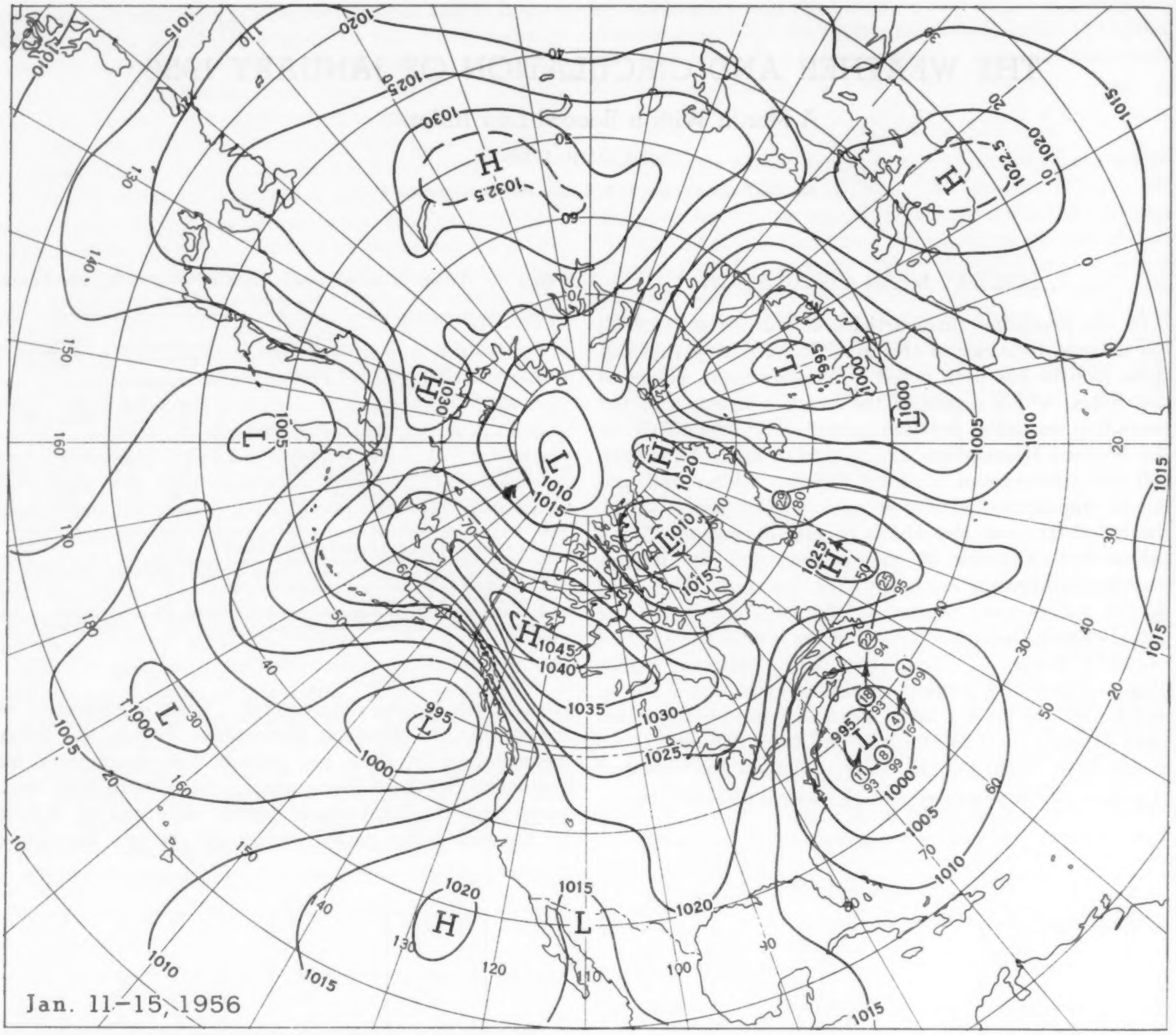

FIGURE 3.—Mean sea level pressure (in mb.) for the 5-day period of lowest zonal index, January 11–15, 1956. The track of the low center off the Middle Atlantic Coast on a series of 5-day mean charts during the month is given by the solid arrows, except dashed in regions of uncertain continuity. The number inside each open square is the last day of the 5-day period, and the number below is the intensity of the center in mb. (last two digits only).

the year from October 1932 through March 1939. The results, shown in table 1, indicate that the index for the total hemisphere was still at a record low level for 700 mb. and a near record at sea level.

In view of the extreme nature of these 5-day mean zonal indices, it is of considerable interest to examine other features of the circulation which accompanied them. The dotted line in figure 2 gives the sea level pressure profile for the Western Hemisphere during the period, January 11–15. A complete reversal from the normal January profile (dashed line) is at once apparent. Average pressures were well below normal at all latitudes from 20° to 45° N., but above normal from 50° to 80° N. In fact, easterlies instead of westerlies prevailed between 35° and 55° N., so that the sea level index was negative.

The 5-day mean isobars at sea level during the period January 11–15 are shown in figure 3. This map contains most of the features customarily associated with low zonal index [5, 6], including split Icelandic and Aleutian Lows, extensive polar anticyclones, weak subtropical Highs, and strong meridional circulation at middle latitudes. Between 35° and 55° N. westerlies were in evi-

dence only in the areas of Europe and the eastern Pacific. In the remainder of the Northern Hemisphere mean winds were from the east in this latitude belt, with the easterlies strongest, and hence the local index lowest, in North America and the central Pacific.

One of the most important contributions to the low zonal index was made by the deep mean Low off the Middle Atlantic Coast of the United States. The location and central intensity of this center of action during each 5-day mean period of January have therefore been superimposed on figure 3. After originating south of the Maritime Provinces at the beginning of the month, the Low apparently moved slowly southwestward and deepened until January 11, then took a northeastward trajectory along the edge of the Gulf Stream until the 25th, and finally amalgamated with the Icelandic Low at the end of the month. Between January 7 and 17 this 5-day mean Low was composed essentially of only one daily cyclone whose trajectory is given in Chart X. It dominated the weather of the eastern United States for over a week, producing gales, high tides, snow, sleet, glaze, rain, and floods in the Northeast and damaging frost in the Southeast, especially in Florida. Further details about this severe quasi-stationary storm can be found in two interesting articles, one by Sable [7] and the other by McQueen and Keith [8].

The 700-mb. circulation was also characterized by marked abnormalities during the period of lowest zonal index at this level, January 7–11. The 5-day mean zonal wind speed profile for the Western Hemisphere, given by the dotted line in figure 4, reveals a minimum wind speed of only 0.5 meters per second between 40° and 45° N., the very latitudes where a westerly maximum of over 14 m. p. s. is normally present in January (dashed line). This marked deficit of wind speed at middle latitudes was associated with two large blocking Highs (around 48° N.), one in mid-Pacific and the other near Newfoundland (fig. 5). Each was accompanied by cut-off Lows to the southwest, difluent areas to the west, and deep troughs both up and downstream. Other typically low index features of figure 5 are the numerous low and high centers, the large amplitude of the waves, and the absence of the normal trough tilt from northeast to southwest.

The abnormalities of figure 5 are brought into sharp focus by the corresponding field of 700-mb. height departure from normal, shown in figure 6. The two blocking Highs were accompanied by the largest anomalies in the Northern Hemisphere, +1,370 feet on the Labrador coast and +1,150 feet in the central Aleutians. Heights at 700 mb. have been above normal in these two areas since the onset of the current spell of low index during the last week of October [1, 2, 3]. In fact, the persistent recurrence of above normal heights in the vicinity of Labrador and Davis Strait was one of the circulation highlights of the year 1955 [9]. The blocking nature of the positive anomalies over Labrador and the Aleutians

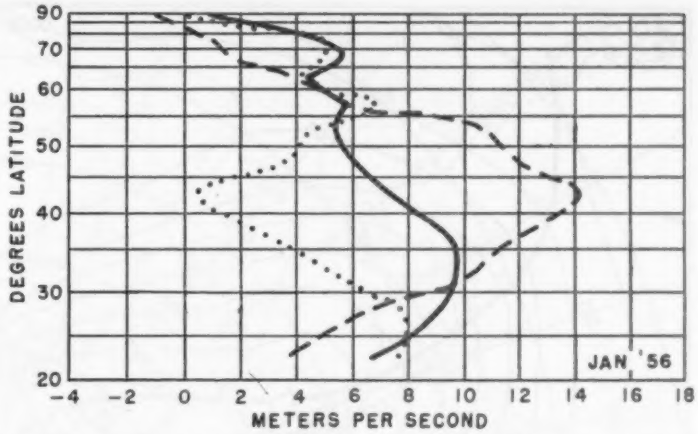

Figure 4.—Mean 700-mb. zonal wind speed profiles in the Western Hemisphere for January 7–11, 1956 (dotted), January 1956 (solid) and January normal (dashed). During both 5- and 30-day periods marked deficit of wind speed at middle latitudes was not fully compensated by small excess at low and high latitudes.

was accentuated by the presence (fig. 6) of pronounced negative anomalies to the south of each center. As a result, strong anomalous flow components from the east at middle latitudes extended from the eastern Atlantic to the Mississippi Valley and also across the central Pacific.

The two largest negative height anomalies in figure 6 were centered off the coasts of the Carolinas (−710 ft.) and Oregon (−570 ft.). The location and central intensity of these two centers on each 5-day mean 700-mb. height departure from normal map during January are plotted in figure 7. The trajectory of the negative center off the east coast was very similar to the track of the 5-day mean sea level Low reproduced in figure 3. Continuity was actually clearer in the former case, however, especially during the first week of the month. In terms of both sea level pressure and 700-mb. height anomaly, the center deepened as it moved southwestward, reaching maximum intensity at the minimum of the 700-mb. index cycle (Jan. 7–11). As the Low migrated northeastward during the remainder of the month, the intensity of its 700-mb. height departure from normal weakened steadily, but its sea level pressure remained sensibly constant.

Figure 7 also contains the tracks of the two largest centers of positive 700-mb. anomaly in figure 6. The magnitude of the Aleutian center increased from the 1st to the 11th of January and then diminished as it moved southward during the last 2 weeks of the month. The center over Labrador also attained its maximum intensity during the 5-day period of lowest 700-mb. index, January 7–11. It was actually formed by an amalgamation of two separate 5-day mean positive anomaly centers, one moving eastward across Canada and the other retrograding from mid-Atlantic. Although the central intensity of both centers increased as they approached Labrador, the intensity of the unified center (+1,370 ft.) was greater than expected from simple extrapolation. The motion

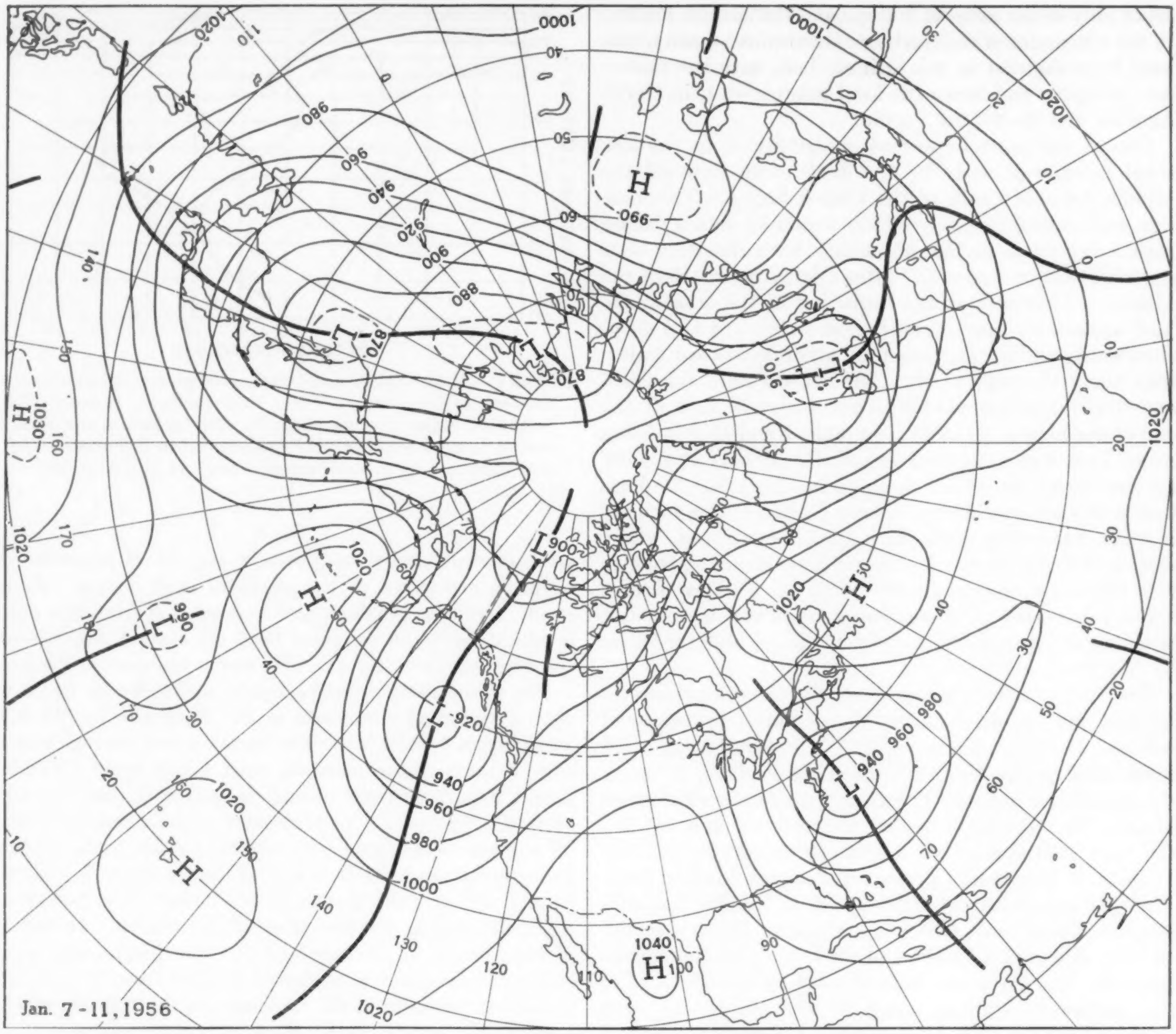

FIGURE 5.—Mean 700-mb. height (in tens of feet) for the 5-day period of lowest 700-mb. index, January 7–11, 1956. Note complete reversal of normal westerly flow between 35° and 55° N. on the east coast of North America.

of the first center was associated with the migration and intensification of a strong daily anticyclone, whose track is reproduced in Chart IX. This High originated in the northeast tip of Siberia on January 3, moved southeastward across Alaska and central Canada, and then curved northeastward to reach Davis Strait on January 9. Here its central intensity of 1064 mb. exceeded by some 25 mb. the highest sea level pressure ever observed in this area [8]. During the next week both the daily anticyclone and the 5-day mean 700-mb. height anomaly weakened and moved southeastward. The daily High then dissipated, but the 5-day mean anomaly was reinforced by a new blocking surge as it moved northwestward during the remainder of the month.

2. MONTHLY MEAN CIRCULATION

When considered on a monthly mean basis, the zonal index was also at a record low level during January 1956, as indicated in table 1. The sea level index, computed for the Western Hemisphere alone, was the lowest for any month of record, while the 700-mb. index was farther below normal than in any previous month. In order to take advantage of sea level indices computed by Willett [10] for the entire Northern Hemisphere from 1899 to 1939, and recently brought up to date in the Extended Forecast Section, the zonal index for January 1956 was recomputed to include the entire hemisphere. The resulting index was 2.5 m. p. s. below the January normal,

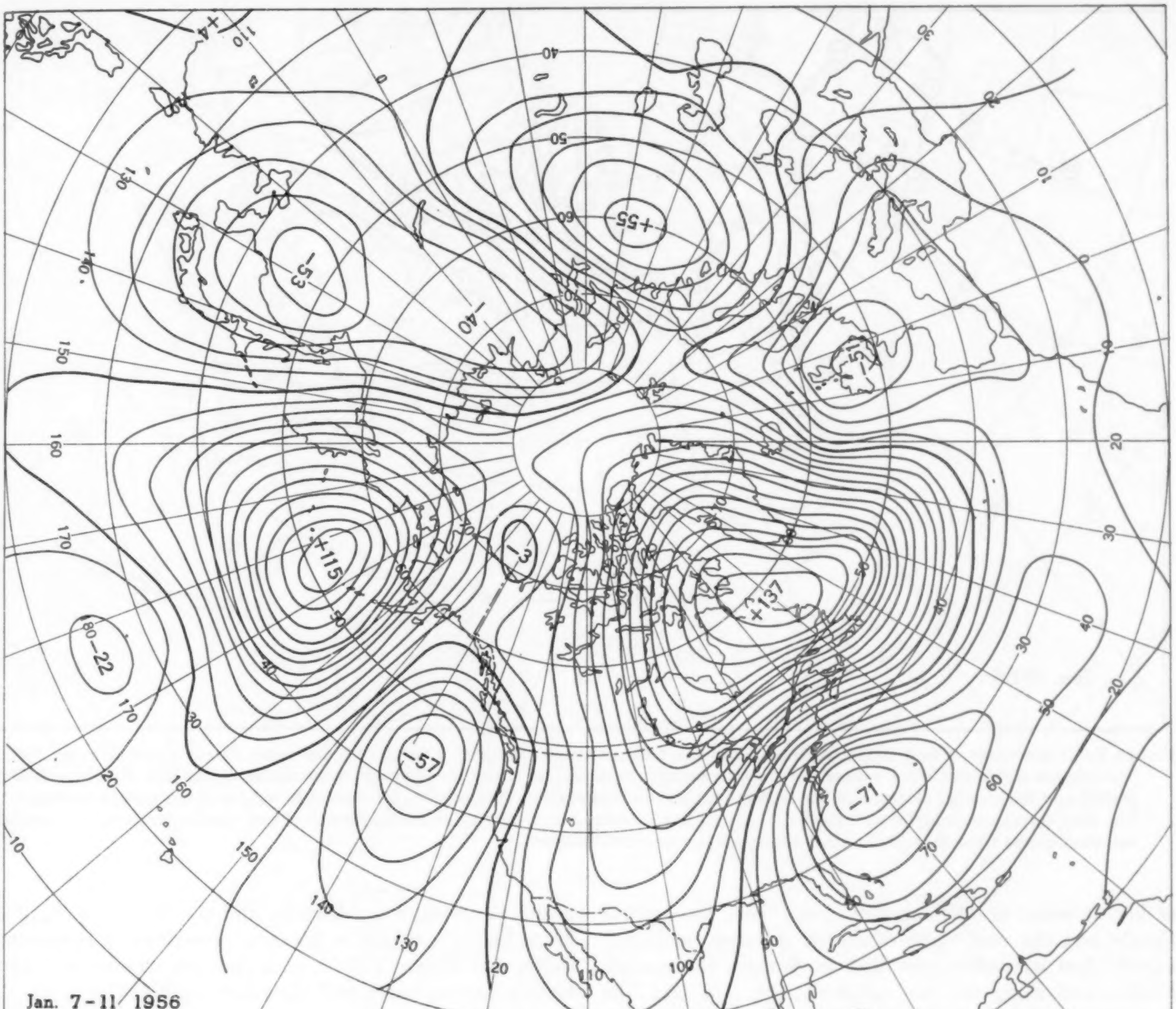

FIGURE 6.—Five-day mean 700-mb. height anomalies (in tens of feet) for the same period shown in figure 5. The anomaly pattern was actually simpler than the contour pattern since subtropical high cells over Mexico, Hawaii, and western Pacific in figure 5 lost their separate identity in this anomaly chart.

the greatest negative departure from normal for any month of a 47-year period of record.

The monthly mean profiles averaged over the Western Hemisphere are drawn as solid lines in figure 2, for sea level pressure, and in figure 4 for 700-mb. wind speed. Both lines fall about half way between the January normal profile (dashed) and the extreme 5-day mean values (dotted). Nevertheless low index characteristics were well-marked for the month as a whole. Pressures were above normal at all latitudes north of 45° N., but below normal to the south. Zonal wind speeds at 700-mb. were weaker than normal between 30° and 55° N., but stronger than normal elsewhere. The west wind maximum was displaced southward by about 10° of latitude and weakened by almost 5 m. p. s.

The mean 700-mb. chart with superimposed height departures from normal for January 1956 (fig. 8) contains most of the features already discussed in connection with the extreme 5-day mean map of January 7–11 (figs. 5 and 6). For example, the greatest departures from normal in figure 8 are the positive anomalies near Labrador and the Aleutians and the negative anomalies off the Carolina and Oregon coasts. These centers were remarkably persistent during the month, and each could be identified during every 5-day mean period, as indicated by figure 7. As on the 5-day mean map, the areas

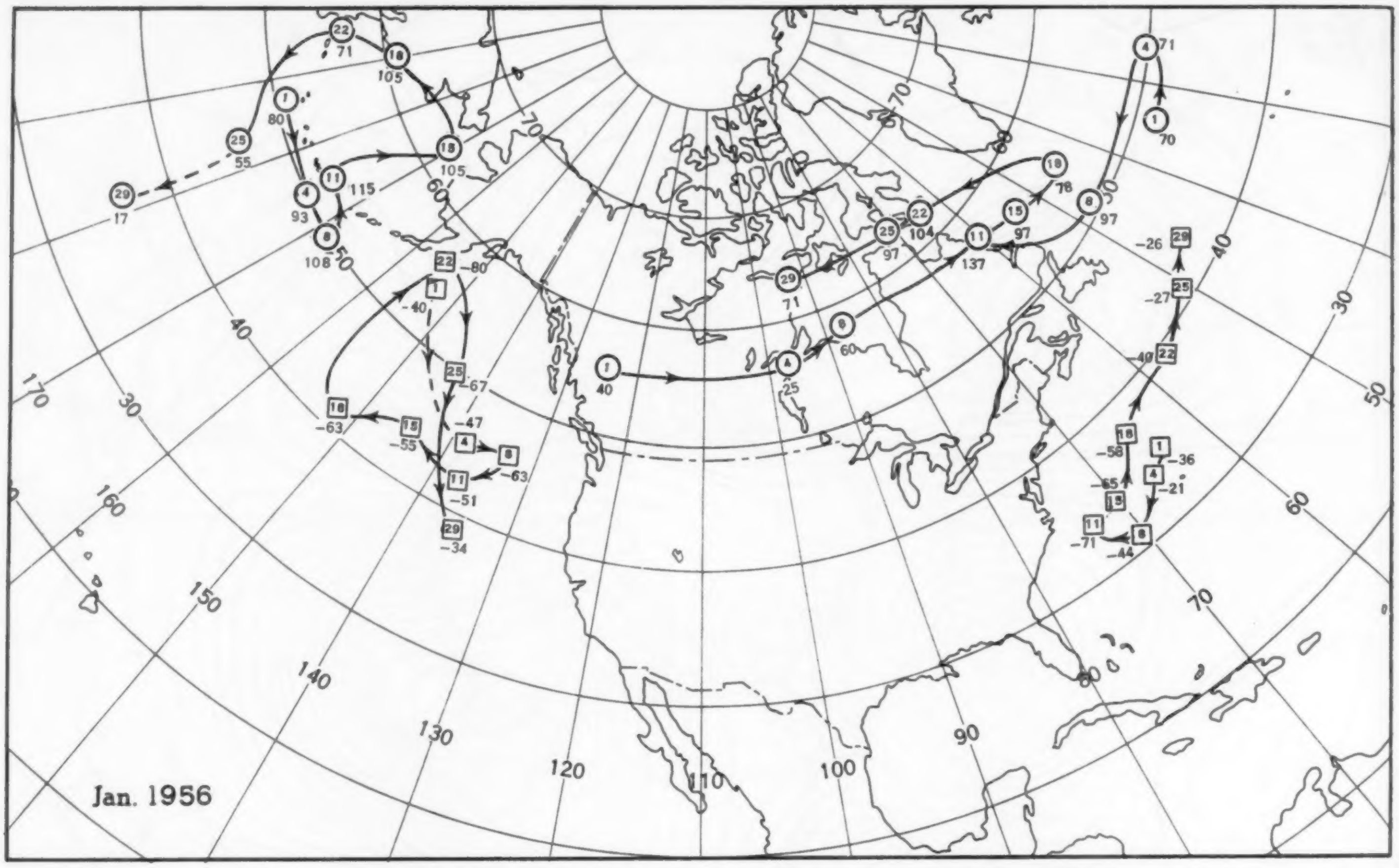

FIGURE 7.—Trajectories of four largest centers of 700-mb. height anomaly in figure 6 on all of 5-day mean charts during January 1956. The number inside each open square (for the two negative centers) and circle (for the positive anomalies) is the last day of the 5-day period, and the number alongside is the central intensity in tens of feet. Trajectories are dashed in regions of less certain continuity. The most intense anomaly of the month (+1,370 ft. in Labrador, Jan. 7–11) was an amalgamation of two separate centers, one moving eastward across Canada, and the other retrograding from mid-Atlantic.

of lowest index on the monthly mean were the central Pacific and the east coast of North America, while relatively fast westerlies prevailed in Europe, the eastern Pacific, and along the east coast of Asia. In fact, the January zonal index (35°–55° N.) was slightly above normal when computed for the Eastern Hemisphere alone.

This regional differentiation of wind speed is well illustrated by figure 9, which gives the geographical distribution of geostrophic wind speed computed from the monthly mean 700-mb. chart (fig. 8). Centers of slow wind speed in Maine and the central Pacific were close to the axis of maximum wind speed on the normal January map (dashed line in fig. 9A). As a result this month's wind speeds were as much as 16 m. p. s. below normal in these two areas (fig. 9B). On the other hand, wind speeds averaged above normal in Europe, the eastern Pacific, and the Asiatic coast. Perhaps the most noteworthy feature of figure 9A is the fact that the solid arrow delineating the principal axis of maximum wind speed was displaced south of its normal position in all parts of the hemisphere from Lake Baikal eastward to the mid-Atlantic. However, secondary axes of maximum wind speed were present at high latitudes, north of the positive anomalies in Labrador and the Aleutians (fig. 8). Thus, blocking existed in the sense prescribed by Berggren, Bolin, and Rossby [11]; i. e., a split jet stream with one branch passing north and the other south of the block.

3. TRANSITION DURING THE MONTH

Figure 10A presents the 15-day mean circulation pattern at 700 mb. over the Northern Hemisphere during the first half of January 1956. Both the contours (solid) and height departures from normal (dashed) closely resemble their 5-day mean counterparts of January 7–11 (figs. 5, 6). By the second half of the month, however, several interesting changes had occurred, as illustrated in figure 10B. In the Eastern Hemisphere an incipient blocking High developed just north of Novaya Zemlya. This was accompanied by marked deepening of a trough in eastern Europe. The resulting easterly anomalous flow in northern Europe was the prelude to the severe weather which overspread the entire continent during February.

In the Western Hemisphere the two strong positive anomaly centers between 50° and 60° N. weakened and

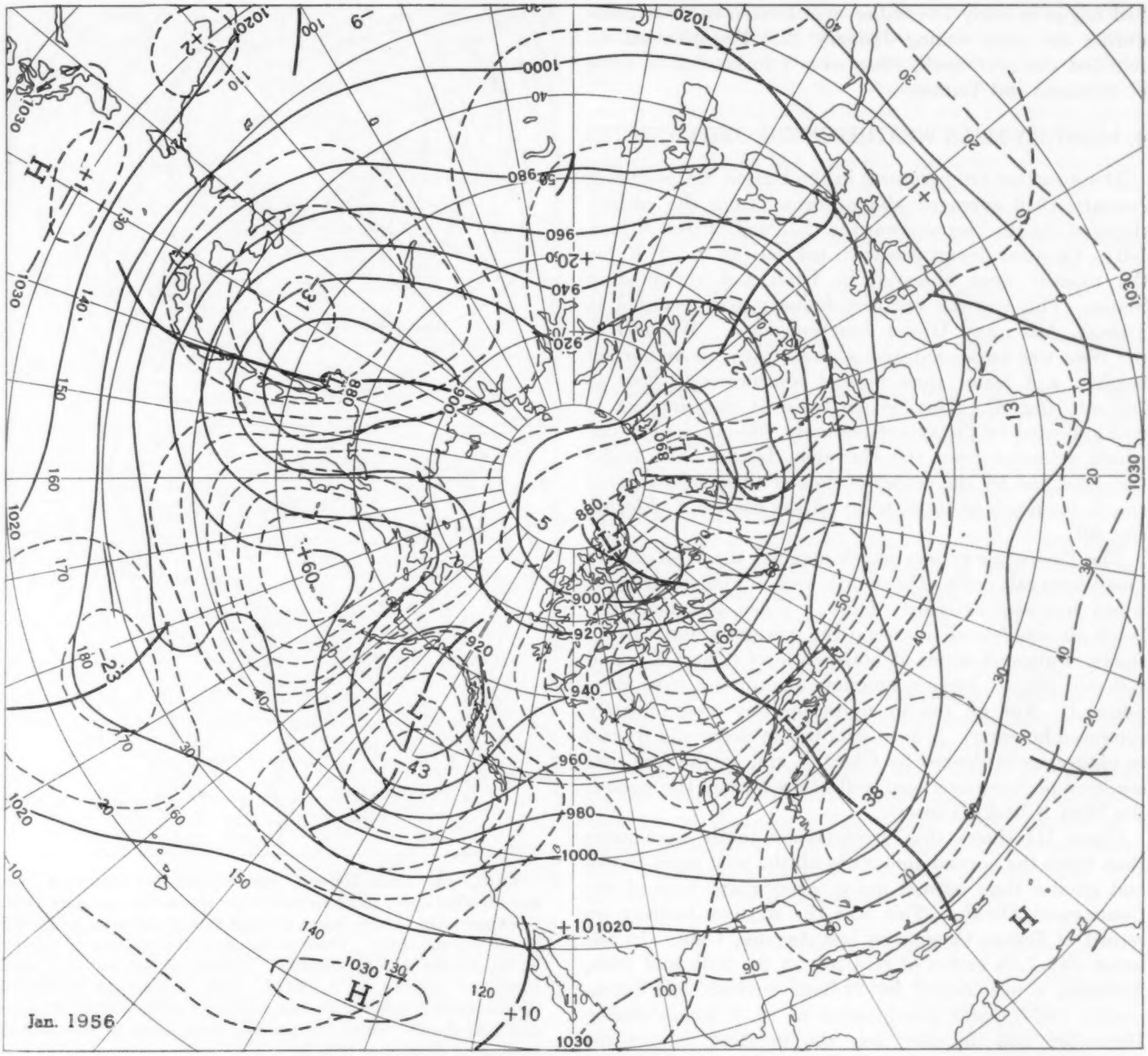

Figure 8.—Thirty-day mean 700-mb. height contours and departures from normal (both in tens of feet) for January 1956. Outstanding features were large centers of positive anomaly in eastern Canada and Aleutians, surrounded almost completely by negative anomalies.

retrograded from the first to the second half of January. The trough in the eastern Pacific also retrograded; but the trough off the east coast moved eastward as the flow pattern in the Atlantic flattened, so that the trough was no longer blocked by a strong ridge in mid-ocean. At the same time that the wavelength across the United States thus increased, the speed of the westerlies in North America diminished as retrogression of the Atlantic block produced a closed High over Hudson Bay. The result was formation of a new mean trough in the familiar manner [12]. The new trough was located in the central United States, where above normal heights during the first half of the month were replaced by below normal heights in the second half.

The effect of these circulation changes upon the temperature regime in the United States was marked. Following generally warm conditions during the first week of January, progressive, westward cooling occurred, with coldest weather during the second week of the month in the Southeast, the third week in the Great Plains, and the fourth week in the Northwest. The changes in precipitation were even more dramatic. In the first half of the month little or no precipitation fell in most southern and central portions of the Nation. This dry spell, which

had begun in early December, was broken in most areas during the week ending January 22. Precipitation intensified the next week, when over 4 inches fell in parts of Arkansas and Tennessee.[2]

4. MONTHLY MEAN WEATHER IN THE UNITED STATES

Mean surface temperatures in the United States during January 1956 averaged above normal from the eastern slopes of the Rockies westward to the Pacific Coast (Chart I-B). Greatest departures from normal (up to 10° F. for the month) were observed in Utah and surrounding States. This was the warmest January ever recorded at Tucson, Ariz., and Grand Junction, Colo. Warmth in the West was associated with a mean ridge, above normal heights, and faster than normal southwesterly flow at 700 mb. (fig. 8). Domination by mild maritime air is well indicated by the axis of the mean 700-mb. jet stream, which extended from the Hawaiian Islands into Idaho (fig. 9A), and by the center of greater than normal wind speeds (as much as 10 m. p. s.) off the coast of California (fig. 9B).

The Pacific air masses which invaded the West at frequent intervals during the month were not only extremely warm but also extremely moist. When this moisture-laden air was forced to ascend the western mountains by strong southwest winds at both sea level (Chart XI) and 700 mb. (fig. 8), copious amounts of rain and snow were released. Not all the precipitation was orographically produced however. A good deal was cyclonic and frontal in nature, as evidenced by Chart X and by the fact that sea level pressure averaged well below normal throughout the West (Chart XI inset).

Chart III shows that precipitation totals were more than twice the normal amount in all the west coast States and greater than normal nearly everywhere west of the Continental Divide. This was the wettest January on record at Burns, Oreg., and Los Angeles, Calif. In the latter city 7.18 inches of rain fell on the 25th and 26th, including a new record for 24-hour amount. At Sacramento, Calif., total precipitation of 19.78 inches during December and January was the greatest consecutive 2-month total since 1862. In addition, January 1956 was the cloudiest month on record at Burbank, Calif. and Yuma, Ariz. (Chart VI). Although floods were widespread during the month, they were not as disastrous as those of December [3].

Like the west coast, the North Atlantic States were under the influence of extremely moist and mild air masses from the adjacent ocean during most of the month. In this case, strong anomalous flow components from the east at both sea level (Chart XI inset) and 700 mb. (fig. 8) carried air masses with a long trajectory over the Atlantic Ocean into this region. The result was above normal values of both temperature (Chart I-B) and precipitation (Chart III). This was the warmest January on record at Caribou, Maine, and the wettest at Blue Hill, Mass. Onshore cyclonic flow was most extreme from January 7 to 17. (See sec. 1 and figs. 3, 5, 6.) During this period Boston, Mass. recorded 5.70 inches of precipitation and 10 consecutive days without sunshine.

Easterly anomalous flow was also responsible for bringing some mild Atlantic air and above normal temperatures to the western Great Lakes region and upper Mississippi Valley. In this area the Lakes themselves probably acted as a local heat source. This is indicated by the fact that the greatest departures from normal occurred on the western shores of the Lakes (+6° F. in Duluth, Minn., +4° F. in Milwaukee, Wis.), while temperatures

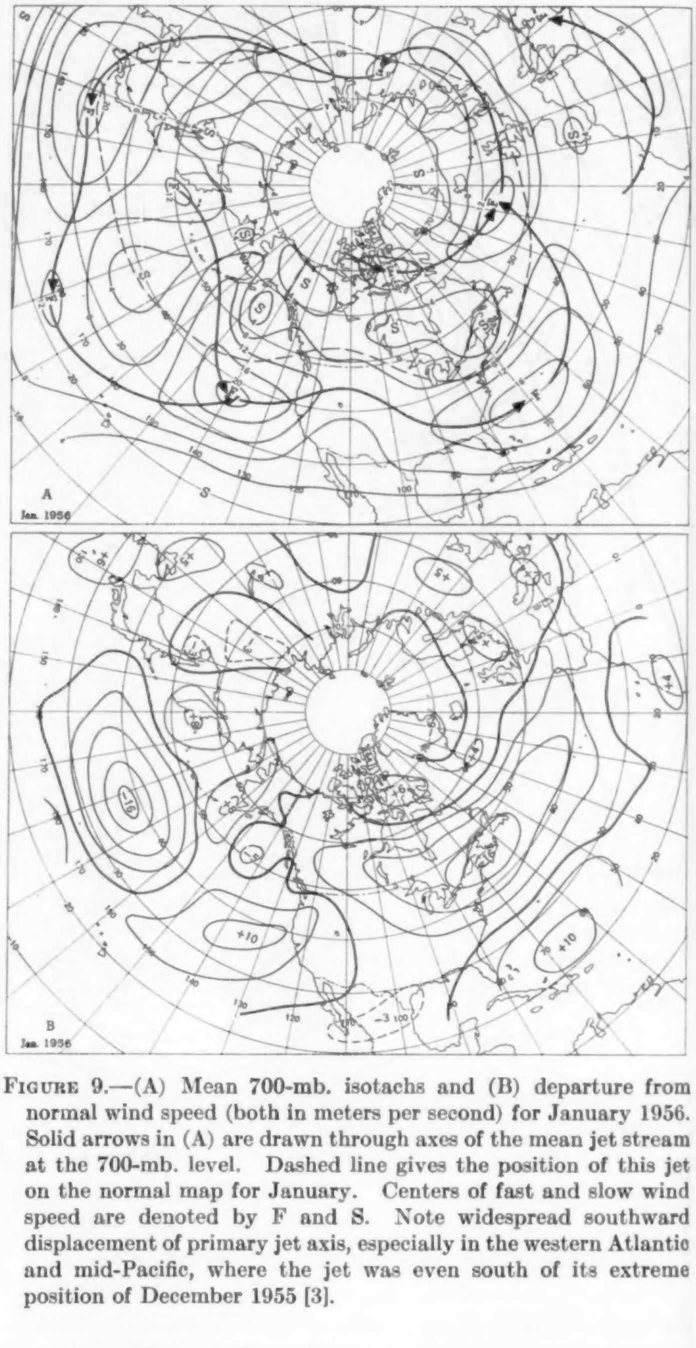

FIGURE 9.—(A) Mean 700-mb. isotachs and (B) departure from normal wind speed (both in meters per second) for January 1956. Solid arrows in (A) are drawn through axes of the mean jet stream at the 700-mb. level. Dashed line gives the position of this jet on the normal map for January. Centers of fast and slow wind speed are denoted by F and S. Note widespread southward displacement of primary jet axis, especially in the western Atlantic and mid-Pacific, where the jet was even south of its extreme position of December 1955 [3].

[2] Charts showing the weekly patterns of temperature and precipitation are given in the *Weekly Weather and Crop Bulletin, National Summary*, vol. XLIII, Nos. 2-5, January, 1956.

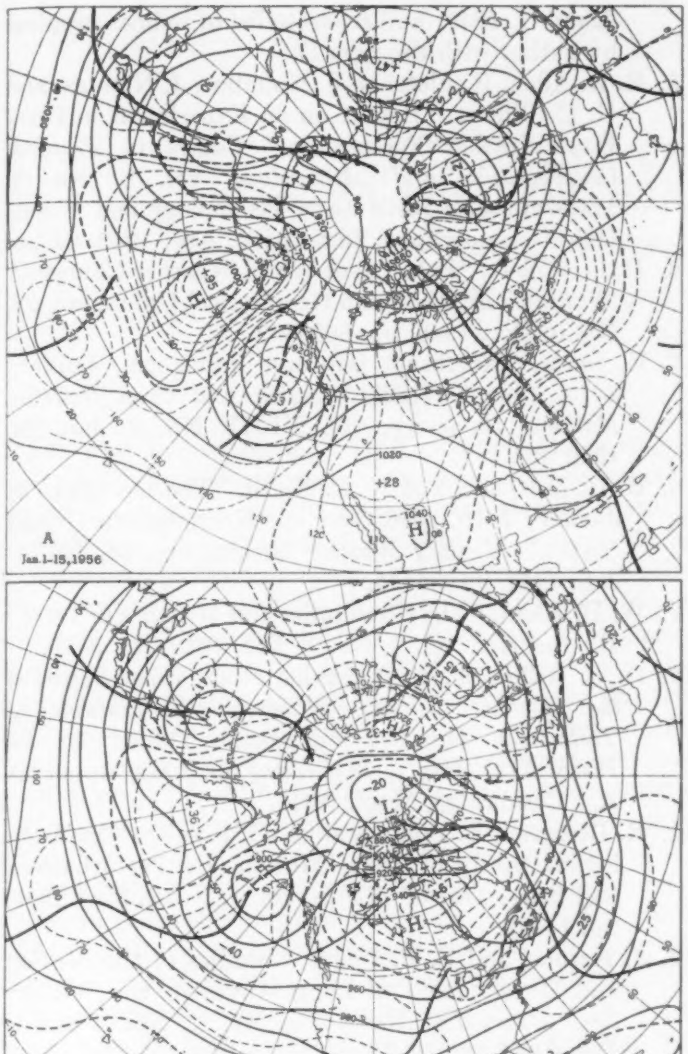

FIGURE 10.—Fifteen-day mean 700-mb. height contours and departures from normal (both in tens of feet) for periods (A) January 1–15, 1956 and (B) January 17–31, 1956. Note formation of new trough east of Continental Divide during second half of month as wave length increased between progressive trough in western Atlantic and retrogressive trough in eastern Pacific, while the westerlies in North America were weakened and displaced southward by development of blocking High over Hudson Bay.

were slightly below normal over the easternmost Lake (Lake Ontario). Precipitation in the Great Lakes region was generally subnormal. Chicago, Ill., reported its second driest January in 86 years of record, while Sault Ste. Marie, Mich. enjoyed its sunniest January in history (Chart VII). This fair weather was a consequence of a stronger than normal mean ridge through the western parts of Hudson Bay and the Great Lakes at both sea level and 700 mb. The large number of daily anticyclone tracks north and west of the Great Lakes should also be noted (Chart IX).

Although easterly anomalous flow brought mild weather to New England and the western Lakes, it was accompanied by below normal temperatures in the northern Plains. Air reaching this area was cooled by passage over snow-covered land (Charts IV and V) and by ascent up the eastern slopes of the Rockies. Foehn warming was minimized by the prevailing easterlies as many daily anticyclones passed north of the area (Chart IX) and many daily cyclones were centered to the south (Chart X). More than twice the normal amount of precipitation, nearly all in the form of snow (Chart V-A), fell in North Dakota, where a minor trough was present at 700 mb. and the mean flow at sea level was cyclonically curved.

In the remainder of the United States, everywhere south of about 40° N. and east of about 95° W. both temperature and precipitation were generally below seasonal normals. The greatest temperature departures occurred in the South Atlantic States. This was the coldest January on record at Orlando, Fla., and the second coldest at Miami. New daily minimum temperature extremes were established during the month at the latter city and also at Savannah, Ga., and West Palm Beach, Fla. The deficiency of precipitation was even more marked. It was the driest January ever observed at Lynchburg, Va., Greensboro, N. C., and Burlington, Iowa. St. Louis, Mo., reported 46 consecutive days without measurable precipitation from December 2, 1955 to January 17, 1956, by far the longest such period of record. Cold dry weather in the southeastern quarter of the Nation was the result of northwesterly flow at 700 mb. between a ridge over the Rockies and a trough off the east coast, the ideal conditions for light precipitation in the Tennessee Valley [13]. The presence of a strong mean ridge at sea level and northerly anomalous flow components at both sea level and 700 mb. were additional contributing factors.

5. PERSISTENCE FROM DECEMBER

The large-scale features of the January 1956 circulation pattern (fig. 8) were extremely persistent from those of the preceding December [3], except over Eurasia. Both months were characterized by deeper than normal troughs off the east coast of the United States, in the eastern Pacific, and near Midway, and by stronger than normal ridges in the Bering Sea, Hudson Bay, and Greenland. A numerical measure of this persistence is given in the first line of table 2, which expresses the correlation coefficient between December and January anomalies of monthly mean 700-mb. height at standard intersections from 30° to 50° N. and 70° to 130 ° W. This year's persistence correlation was considerably higher than expected from random data and slightly higher than given by Namias [14] for the years 1942–50. It is interesting to note that Namias obtained a higher correlation coefficient between December and January than between any other pair of months.

TABLE 2.—*Persistence of monthly mean anomalies in the United States from December to January*

	1955–56	Normal (1942–50)	Random
700-mb. height (correlation)	0.42	0.39	0
Temperature (0 or 1 class change, %)	82	69	59
Precipitation (0 class change, %)	48	37	33

Persistence of the weather elements in the United States from December 1955 to January 1956 was even more striking, as shown by the last two lines of table 2. Of 100 stations evenly distributed over the country, there were 82 in which the temperature anomaly did not change by more than one class (out of five) and 48 in which the precipitation remained in the same class (out of three). This represents considerably greater persistence than expected either by chance or from past years [14]. Furthermore, extreme changes from this December to January were experienced by only 3 stations in temperature (3 or 4 class change) and 9 stations in precipitation (2 class change). It is possible that this year's unusual persistence was related to the prolonged period of low zonal index since Namias [14] has suggested that persistence is greater at times of low index.

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THE ICE STORM OF JANUARY 7-10, 1956 OVER THE NORTHEASTERN UNITED STATES

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1. INTRODUCTION

The first major storm of 1956 affecting the northern Atlantic Coastal States was an intense one of the "northeaster" type which caused an extensive area of freezing precipitation. On January 7, the weather patterns of the storm were developing both at the surface and aloft over the Atlantic Ocean and North America. A strong block south of Greenland and an intense Low east of Hatteras combined to produce an area of glaze which began as sleet and freezing rain shortly before midnight on the 7th in eastern Maine and spread rapidly westward and southward. The southern limit of these hydrometeors was reached on the 10th in the Carolinas and Tennessee with the westernmost limit occurring in Michigan on the same date. The duration at any one station was quite variable, ranging from 5 to 15 hours in the coastal States to over 75 hours (intermittently) in portions of Ohio. Only the rapid transition of temperatures from below freezing to thawing conditions kept this storm from becoming one with major ice damage. However, from onset to the very end, these freezing types of precipitation persisted for seven days over some portion of the northeastern States. In this article we shall for the most part confine our discussion to the first 4 days.

2. ANTECEDENT CONDITIONS

During the 5 weeks preceding this storm, the storm tracks over the eastern portion of the nation were primarily along the northern border of the country. Under such conditions, precipitation totals had been small and generally near or below normal temperatures had prevailed. (See Andrews [1].)

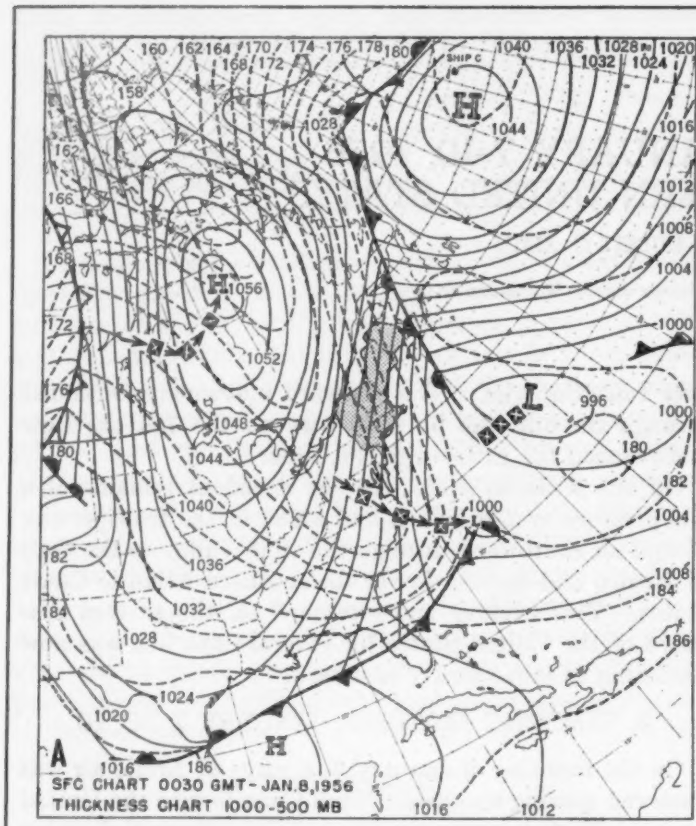
The low pressure system which was partially responsible for the ice storm had its inception along the Pacific polar front on January 4 when it first appeared in association with a short-wave trough aloft that was moving out of the long-wave position just off the west coast. Advancing along the northern border of the United States just south of the thermal field of an Arctic front, the Low remained weak as it passed through the long-wave ridge position over central United States. A new development occurred at the point of occlusion on the 6th, in the vicinity of Lake Michigan, as the original center weakened and dissipated over Lake Superior. The eastward movement continued through the 6th, and by 0030 GMT on the 7th the Low had reached the Pittsburgh, Pa. area.

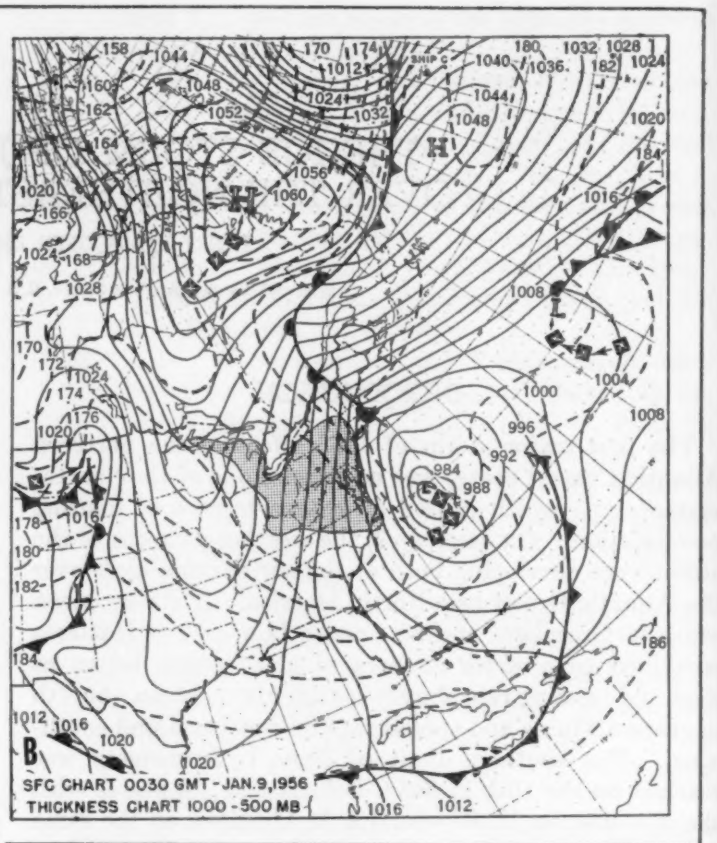
Off the Atlantic shore was the decadent remains of a low pressure system which had aided in the transporting of cold air southward through the Ohio Valley to the Gulf of Mexico and eastward over the southern Atlantic Coast States. This cyclone had remained off the eastern seaboard of the United States for several days but was now beginning to move slowly eastward.

3. SYNOPTIC FEATURES, JANUARY 7 TO 10

On the morning of January 7, a surge of Arctic air was observed moving southward from Canada into the United States. With pressures building and gradients tightening to the west and north of the cyclone located near Pittsburgh, much of the continent from north of Hudson Bay southward to the Gulf of Mexico was under the domination of a large anticyclone centered just south of Churchill, Manitoba. In the upper air a building ridge extended north-northeastward from Texas to Hudson Bay. As the dominant surface anticyclone and ridge aloft turned the winds to a more northerly direction, the Pittsburgh Low, as well as the associated cut-off Low aloft near Buffalo, was now changing its course to a more southeasterly direction. At the same time a well-developed blocking High pressure cell was centered near stationary ship "C" (52°45′ N., 35°30′ W.), but its retrogression toward Newfoundland was clearly indicated by intense deepening in the Iceland to British Isles area.

The strong confluent northwestward flow of marine air between the pressure systems in the Atlantic had carried the warm front to the north of the now-dying cyclone. Nevertheless, the front remained well-defined, with the temperature differential across it very sharp both at the surface and aloft. The maritime provinces of Nova Scotia and Newfoundland reported temperatures in the upper 40's while readings on the continent were in the 10° F. to 20° F. range. Aloft the 1000–500-mb. thickness for 0300 GMT of the 7th indicated a strong thermal gradient west of this front, approximately 700 feet in 180 miles. This gradient would indicate thermal winds of about 100 knots, in close agreement with the 110-knot thermal wind computed from the Quebec rawin.





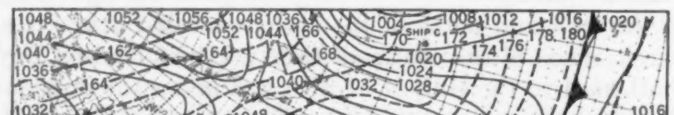


FIGURE 1.—Sea level and 1000–500-mb. thickness synoptic patterns for January 8–10, 1956. Shaded area represents region in which freezing rain or sleet occurred in the 24-hour period 0001 to 2400 EST of each date. (A) 0030 GMT surface chart for January 8, 1956 with 1000–500-mb. thickness for 0300 GMT in dashed lines at 200-foot intervals. (B) 0030 GMT surface chart for January 9, 1956 with 1000–500-mb. thickness for 0300 GMT. (C) 0030 GMT surface chart for January 10, 1956 with 1000–500-mb. thickness for 0300 GMT.

By 0030 GMT of January 8, the Pittsburgh Low had crossed the coast in the vicinity of Cape Hatteras and was located at a point 300 nautical miles east-southeast of that area (fig. 1A). The 500-mb. low center of 17,500 feet was over the Hatteras region (fig. 2A). Simultaneously, the blocking conditions south of Greenland had begun to retrograde. The relative movement of these two systems served to produce a general tightening of the pressure gradient along the New England and Nova Scotia coasts. Thus, with the increased easterly wind flow the stationary front off the coast began moving westward as a warm front. Shortly after the 0030 GMT synoptic chart freezing rain and sleet began over the southeastern half of Maine. Over the interior of Canada the surface high pressure center moved into the eastern portion of Hudson Bay and continued to intensify. And at the same time the 500-mb. ridge built northward with the 18,200-foot

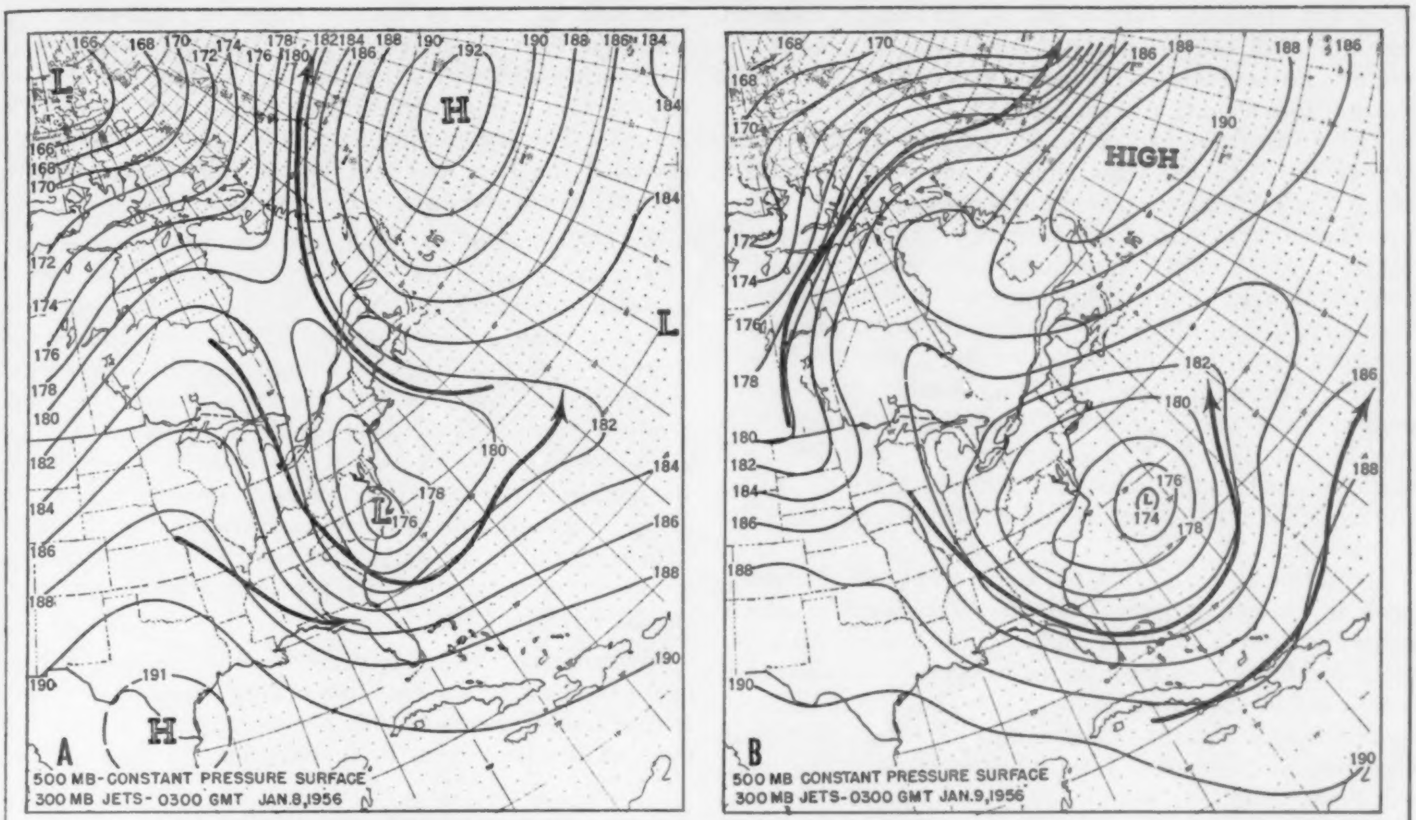


FIGURE 2.—500-mb. constant pressure synoptic patterns with 300-mb. jet stream, 0300 GMT. (A) January 8, (B) January 9, (C) January 10, 1956.

contour reaching into the southwestern portion of Hudson Bay.

During the next 24 hours the high pressure block, with a center apparently south of Greenland, assumed an oval shape as its western periphery extended into Labrador, and the eastern limits were over the Azores. Concurrently, the continental surface High and upper ridge extended across Hudson Bay and northern Quebec. The surface center continued to build reaching an extreme value of 1064 mb. over northern Quebec. By now the cyclone had moved into the long-wave trough both at the surface and aloft and had begun to recurve northward (figs. 1B and 2B). The central pressure of the Low continued to fall with the current reading under 980 mb. at the surface and 17,400 feet at 500 mb. The surface and upper centers assumed a more nearly vertical position and the 300-mb. jet had symmetrically encircled the southern half of the Low with winds in excess of 100 knots over much of the distance. Pressure gradients continued to intensify, becoming extreme east of the warm front and producing wind gusts of 45 to 60 knots along the coastal area. The increased east-west flow aloft north of the Low center

carried the ice shield farther inland. It had, by midnight EST of the 8th, covered part of New Jersey, extreme northeastern Pennsylvania, and more than half of New York State. (See shaded area on fig. 1B.)

The warm front had continued moving westward and was by then crossing Massachusetts and New Hampshire (fig. 3). Following the passage of the warm front, temperatures rose rapidly, sleet and freezing rain ended, but rain continued. The rapid transition from freezing to thawing conditions, even during night hours, was well illustrated at Danbury, Conn., where at 6 p. m. EST of the 8th the temperature was 15° F. and by 9:30 a. m. of the 9th it was 43° F. The duration of the icing conditions over the coastal States varied generally from 5 to 15 hours. The post-frontal rainfall, it is thought, was in part due to a weak tropical warm front that was overlying the area. Further discussion on this subject will appear later in this article.

By 0030 GMT of the 10th the warm front had reached a position extending from Delaware Bay northwestward across western New York and on into Canada (fig. 3). Sleet and freezing rain continued to spread westward and southward in the strong easterly flow, the limits reached by midnight EST of the 9th extending into northern North Carolina, northeastern Kentucky, western Ohio, and Michigan. Light to locally heavy amounts of rain continued to fall east of the warm front as temperatures remained in the upper 30's or in the 40's. The synoptic situation responsible for these conditions had changed but slightly during the past 24 hours (fig. 1C) with the cold core Low having become vertical and with little change in intensity as it slowly migrated northward. At the same time the northern high pressures combined at both surface and aloft, extending the block to the Labrador region, with a ridge persisting southeastward toward the Azores. The 300-mb. jet (fig. 2C) by now had completely encircled the Low and was coming inland over eastern Massachusetts and on into Pennsylvania.

By 0030 GMT of the 11th, the pressure systems were deteriorating although blocking conditions continued. Advection of warm air both from the east and the west over the Great Lakes region (figs. 1B and C) decreased the thermal gradients finally to a point of frontolysis of the warm front. Nevertheless, overrunning continued to produce freezing rain or sleet for the next 2 days in Ohio. Sleet and freezing rain were reported in some sections of the northeastern States into the 16th, but with decreasing area. In general, the pattern was reversed after the 11th, with cooler drier air moving in from the northwest and gradually forcing the sleet and glaze conditions back across the New England States.

Much of this sleet and freezing rain occurred in a portion of the region where these types of hydrometeors are most frequent. Furthermore, as mentioned by Brooks [2], most of the ice storms in the Northeast fall into three

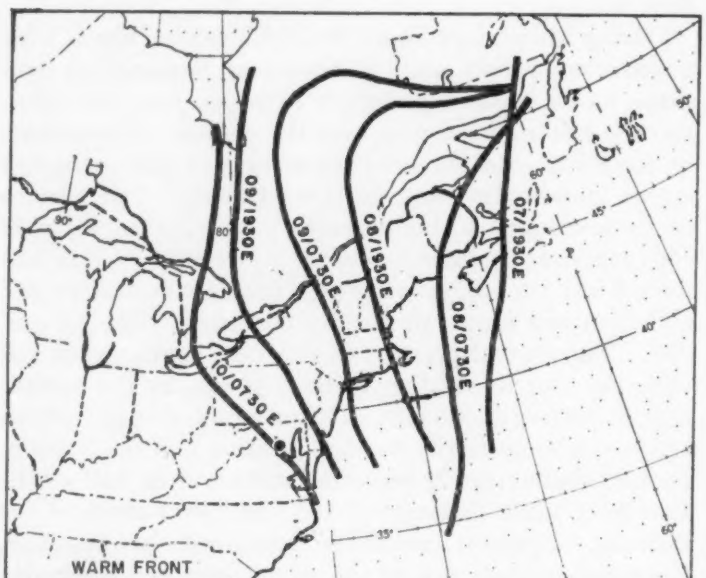


FIGURE 3.—Positions of the warm front at 12-hour intervals during the period 0030 GMT, January 8 to 1230 GMT, January 10, 1956. Times on fronts are in EST for ease in comparison with precipitation. The unusual westward movement of the warm front is clearly depicted on this chart.

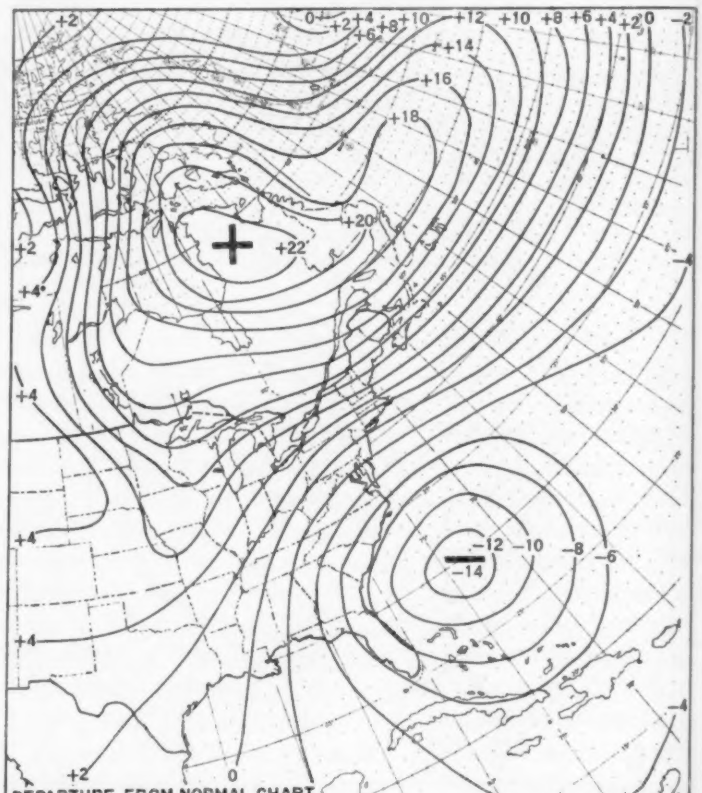


FIGURE 4.—Strength of blocking action in the Atlantic is indicated by the departures of the 500-mb. heights from normal for 0300 GMT, January 9, 1956. Values are in hundreds of geopotential feet.

classifications: (1) warm air arriving over residual cold air, (2) cold air coming in below and warm air arriving above, and (3) cold air pushing in from the north or west below a raincloud. During the period of this ice storm type 1 prevailed through January 11, and then changed to type 3 as the precipitation moved eastward again. The third type is in reality the reversal of the first two, i. e., the forms of precipitation change from rain or freezing rain to sleet and then into snow as the wedge of cold air moves into the area beneath the overcast.

4. BLOCKING ACTION OVER THE NORTHWEST ATLANTIC OCEAN

One of the most important broad-scale features contributing to the prolonged and widespread areas of sleet, freezing rain, and rain was the blocking action of the 500-mb. High in the northwestern Atlantic Ocean. The outstanding effect of this block was that it forced the cyclone along the Atlantic Coast to remain in the long-wave trough for a period of nearly seven days. This produced a strong easterly flow of moist air with a duration period sufficient to permit it to extend into Michigan and Ohio in the west and to the Carolinas and Tennessee to the south.

The strength of this block may have been due partially to the advection of warm air into the Quebec and Labrador region. There were indications during this period that tropical air was transported aloft over the easternmost New England States and portions of eastern Canada. Godson [3] has written that temperatures of −17° C. at 500 mb. are evidence of tropical air in Canada during the winter season, while Vederman [4] delineates the northern limit of tropical air at that level along the −20° C. isotherm. Temperatures at the 500-mb. level for a few of the eastern stations are shown in table 1.

5. THICKNESS PATTERN

Further indication of the extent of the warming from this storm over the Northeast is obtained from the change in thickness of the 1000–500-mb. layer (figs. 1A–C). It is generally known that a 200-foot increase in the 1000–500-mb. thickness is equivalent to a 5.4° F. rise in the mean virtual temperature of the layer. In this case, in northern New York State thickness values increased from 17,100 feet on the 8th to 18,000 feet on the 10th, and in northern Quebec the change was even greater, being from 16,400 feet to 17,600 feet. Thus, there was a thickness change of 900 and 1,200 feet, respectively, or a rise of the mean virtual temperature of the column of over 24° and 32° F., respectively; this is in fair agreement with actual surface temperature changes occurring during this warming period.

The thickness patterns over the eastern States and Canada also furnish an excellent picture of the intensity of the warm front.[1] On the morning of the 8th, for instance (fig. 1A), a tight thermal gradient is shown ahead of the warm front, with thickness values of 17,100 to 18,100 feet between Buffalo, N. Y., and Halifax, Nova Scotia, indicating a front of strong intensity. The easterly flow suggested by the 500-mb. contours (fig. 2A) indicates rapid advection of warm air over New Brunswick, Nova Scotia, and New England. By 0030 GMT of January 9 (fig. 1B) the range of thickness values was only between 17,500 and 17,900 feet, indicating a front of only moderate intensity. Twenty-four hours later (fig. 1C) the thermal gradient was almost destroyed (weak classification), and warm air had by this time been transported into the Carolinas, Indiana, Michigan, and the Hudson Bay region.

TABLE 1.—*500-mb. temperatures (°C.) at selected stations in North America for Jan. 7–10, 1956*

Station	Date and time							
	7th		8th		9th		10th	
	03	15	03	15	03	15	03	15 GMT
Moosonee, Ontario	−32	−33	−28	−30	−18	−13	−24	−19
Nitchequon, Quebec	−29	−32	−29	−21	−18	−20	....	−27
Stephenville, Newfoundland	−17	−17	−17	−16	−23	−25	−20	−19
Sable Island	−19	−14	−16	−18	....	−27	....	−20
Caribou, Maine	−25	−23	−16	−18	−20	−22	−18	−18
Portland, Maine	−26	−24	−19	−17	−21	−20	−17	−21
Nantucket, Mass	−23	−22	−21	−17	−19	−17	−22	−21
Albany, N. Y	−28	−29	−26	−21	−19	−18	−16	−22
Maniwaki, Quebec	−29	−30	−30	−20	−18	−25	−19	−18
Hempstead, N. Y	−25	−26	−24	....	−19	−17	−20	−22
Buffalo, N. Y	−29	−32	−30	−25	−21	−21	−18	−20
Washington, D. C	−23	−30	−31	−26	−22	−18	−18	−20

6. TROPOPAUSE

An interesting comparison of the extreme temperatures in the air surrounding the cyclone and anticyclone during this period is found in the height of the tropopause. As the surface and upper low centers moved off the eastern seaboard on the 7th and 8th in the vicinity of Hatteras, a tropopause value of 425 mb. or a height of 22,000 feet was reported at that station, with a temperature of −38° C. at the tropopause level. This low tropopause height and high temperature were confirmed by comparable values at Norfolk and Greensboro. At 0300 GMT on January 10 in the Goose Bay, Labrador area the tropopause level rose to a height of 42,000 feet (170 mb.) and the temperature fell to −72° C. This height persisted through January 11 at 0300 GMT.

7. DIFFERENTIAL ANALYSIS OF THE SURFACE PRESSURE CENTERS

As the 500-mb. Low was intensifying in the southward plunge during the 7th, the surface low center was moving out ahead of the upper Low and was coming under higher 1000–500-mb. thickness values. On the early morning surface and upper air charts of the 8th, the surface center was under a 500-mb. height of approximately 100 feet

[1] In the National Weather Analysis Center the criterion now used for the determination of the intensity of fronts is the shear across the front of the 1000–500-mb. thermal winds. A difference of 25–49 knots indicates a front of weak intensity, 50–74 knots moderate, and 75 knots or greater strong intensity.

higher than the preceding day, while the thickness value was 400 feet greater. By algebraic subtraction this made a net change of −300 feet at the 1000-mb. level or an indicated deepening of 12 mb. of the sea level Low. From the 8th to the 9th as the cold core of the Low became more nearly vertical, the height lines at 500 mb. lowered by 700 feet above the surface center, and the thickness decreased by 200 feet. Thus a net change of −500 feet was indicated at 1000 mb. or a deepening of the surface Low of 20 mb. That such deepening would occur was also indicated by the 500-mb. chart (fig. 2A). The packing and sharp curvature of contours over the South Carolina coastal region indicated considerable cyclonic vorticity in that region. That this vorticity would be advected eastward was indicated by the geostrophic winds on the mean flow chart for the same time.

In a similar procedure the building of the surface high pressure over eastern Canada can be ascertained. The retrogression of the 500-mb. block to a nearly vertical position over the surface High from the 7th to the 9th inclusive produced a rise of 800 feet at the 500-mb. level over the high center. At the same time the advection of the 1000–500-mb. thickness brought about a rise of 400 feet in thickness value during the same period. Thus there was a net change in the 1000-mb. level of 400 feet or intensification of 16 mb. at mean sea level.

8. DEPARTURES FROM NORMAL

An important aspect of this storm and the blocking high pressure system, aside from its extensive pattern of freezing precipitation and its southward and westward-moving warm front, was the deviation from normal in almost all meteorological categories. At the 500-mb. level the normal January contour gradient between 32° N., 74° W. and 58° N., 68° W. is approximately 2,200 feet (westerly geostrophic winds). At 0300 GMT, January 9 (see fig. 4) the gradient between the anomaly centers was approximately 3,400 feet and by 1500 GMT of that date had increased to about 3,900 feet. In other words, the easterly flow across these 26° of latitude was almost as great as the normal westerly flow. At no time during these 4 days, January 7–10 inclusive, was the gradient between the anomaly centers less than 2,000 feet. For a broad-scale picture of the anomaly pattern, reference may be made to figures 6 and 7 in the preceding article by Klein [5].

The 1000–500-mb. thickness also underwent extreme departures from normal. January 9, 1500 GMT thickness departures (not shown) illustrated this abnormality quite clearly with values 1,400 feet above normal over Quebec, and 600 feet below normal east of Florida, or a total departure of 2,000 feet from the normal between the anomaly centers.

The anomalies of the surface pressure along the eastern seaboard of the North American Continent were likewise outstanding. It suffices to state that the central pressure of the cyclone was the lowest ever recorded in January for that region. From midnight EST of the 9th, through 1230 EST of the 10th, the pressure of the low center varied between 976 and 980 mb. Lennahan [6] in an examination of the Historical Weather Map Series (1899 through 1951) found no January pressures in the same region of less than 981 mb. Simultaneously the anticyclone produced abnormally high surface pressure over northeastern Canada. Lennahan again found no pressures above 1040 mb. for January in the Canadian area, but under this extreme high pressure block the center attained a value of 1064 mb., registering above 1060 mb. from 1200 GMT, January 8 through 1200 GMT of January 9.

R. W. James [7] has computed the frequency of cyclones and anticyclones over North America and their intensities. He found that at latitude 30° to 35° N., the mean central pressure of cyclones was 1008 mb. with a standard deviation of 6.3 mb. The central value of this current storm thus was more than five standard deviations [2] from the mean.

Similarly the mean highest pressure between the latitudes 60° to 65° N., where the highest center was located, was computed to be 1035.5 mb. for winter months, with a standard deviation of 7.5 mb. The actual value at this time was 1064 mb. in the high center. Thus this January's value was computed as nearly four times the standard deviation [2] from the mean.

Another extreme value during this time occurred in the mean westerly flow in the temperate zone extending from 35°–55° N., over the Western Hemisphere. On January 8 the zonal index of the mean flow at 700 mb. was recorded as 1.3 m. p. s., and the 5-day mean from the 7th–11th was the lowest ever recorded, 2.9 m. p. s. For a complete discussion on these anomalies in the westerly flow, see the preceding discussion by Klein [5].

9. STATIONARY ASPECT OF LONG-WAVE TROUGH

By the use of the Hovmöller time-height diagram [8], the positions and persistency of the long-wave troughs, as well as some value of intensity, can be followed. The National Weather Analysis Center prepares this chart every 12 hours for latitudes of 35° and 50° N., extending westward from 10° W. to 140° E. From 1500 GMT, January 7 until 1500 GMT, January 10, the position of the long-wave trough off the Atlantic coast oscillated between 68° and 76° W., while off the Pacific Coast, the long-wave position ranged between 134° and 138° W. The wave length between these east and west coast long-wave troughs varied from 58° to 68° during the period.

The actual stationary wave length at 0300 GMT of the 7th was found to be 53°, a difference from observed of 14°. By 1500 GMT it was 65°, then lowered to 59° through 0300 GMT of the 9th; these four values had a difference from the observed of 1°. At 1500 GMT of the 9th the computed value was 66°; at 0300 GMT of the 10th, 68°; and at 1500 GMT of 10th, 69°; the observed values differed

[2] James' period was for only 4.5 years, thus a sufficient sampling of the population may not have been obtained.

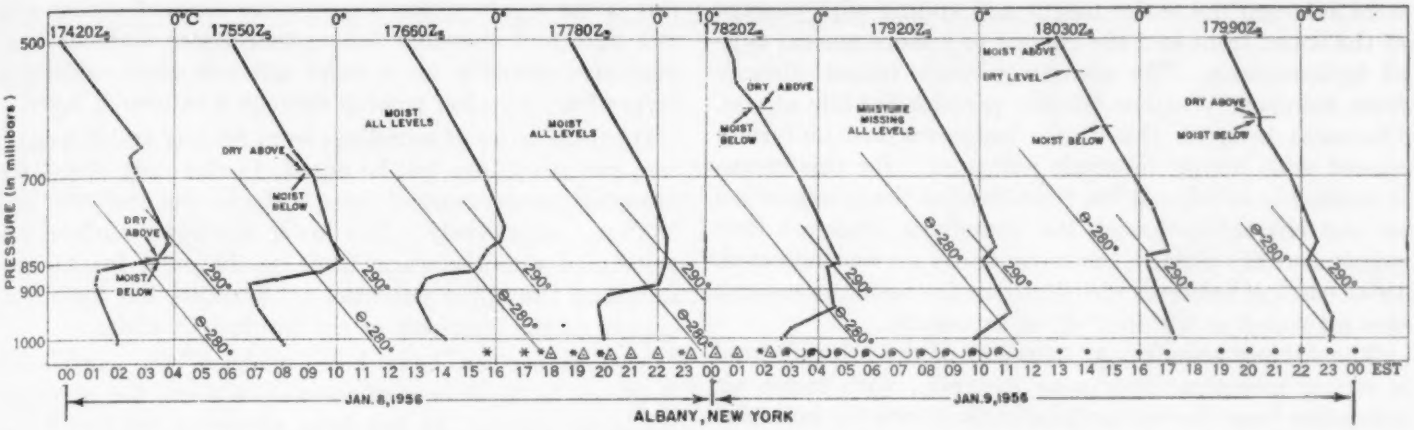


FIGURE 5.—The growth of the above-freezing area of the inversion layer is shown with the resulting change of types of hydrometeors at Albany, N. Y. for January 8–10, 1956. The pseudoadiabatic diagrams are plotted around the 0° isotherms spaced at 6 hour intervals with the hourly type of precipitation indicated beneath the soundings. Times are indicated in EST. 1000–500-mb. thickness values are at the top right of each sounding for comparison with type of precipitation.

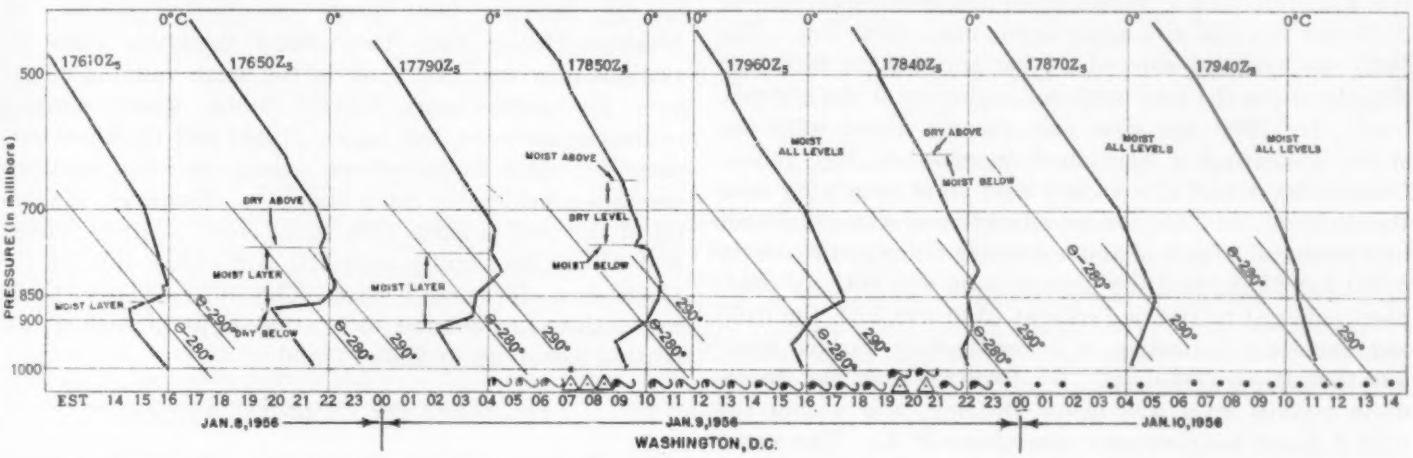


FIGURE 6.—Growth of the above-freezing area of the inversion layer with resulting change in type of hydrometeor, Washington, D. C.' January 8–10, 1956.

by 2°, 1°, 0°, respectively. The formula for wave motion as derived by Rossby [9] and expanded by Cressman [10] was used in computing the stationary wavelength. The close agreement of the computed and observed values of the long-wave troughs indicated that a stable condition existed in the long-wave pattern upstream from the storm center during the period. This fact, plus the strong blocking effect over the northwestern Atlantic, explains why the storm center remained stationary off the eastern seaboard of the United States.

10. PSEUDOADIABATIC CHANGES

The vertical temperature distribution necessary for the formation of sleet or glaze is well known. Temperatures at the surface must be freezing or colder, coincident with an inversion above the surface that extends above the freezing point. Air above the inversion must have a high moisture content with wet bulb temperatures $>0°$ C. so that evaporation of precipitation falling through this air cannot cool it below the freezing point. The initial form of precipitation, i. e., rain or snow, in the upper air matters but little, as the snow will melt in the above-freezing zone if the depth of this zone exceeds 2,000 to 3,000 feet, and warm air is continuing to be advected into the area. The height of the inversion above the surface generally determines the form of the hydrometeor reaching the surface: The greater the distance the more likely that the supercooled rain droplet will fall as sleet. The strength of the low-level winds is also important; strong winds and the resulting turbulence produce conditions more favorable for the formation of sleet. These conditions tend to maintain a deeper layer of cold air as well as creating a stirring action which increases the probability that the supercooled droplets will freeze.

Pseudoadiabatic charts for Albany, N. Y., and Washington, D. C. are shown (figs. 5 and 6, respectively) for each 6-hourly radiosonde, beginning when temperatures

were sub-zero the entire height and ending with passage of the warm front and the change to a more normal type of hydrometeor. The soundings were traced directly from standard Weather Bureau pseudoadiabatic charts. Distances between the 0° C. isotherms are uniformly spaced with hourly intervals indicated. By this means it is possible to indicate the type-trend of the precipitation as the characteristics of the soundings change. Dew points were not plotted, but moist or dry air was indicated. In all cases of moist air identification the moisture content was saturated or within 4° C. of saturation.

The Albany sounding at 0400 EST of the 8th indicated a strong inversion layer near 850 mb., with moist air extending from the surface to slightly above the inversion layer, but with the entire sounding below freezing. During the next 12 hours, warm moist air advection had extended the near-saturation region over the entire sounding below 500 mb. and the upper portion of the inversion was above freezing. Snow flurries first were reported near 0800 EST but did not occur again until 1600 EST. The 1600 EST sounding showed a layer about 2,500 feet deep slightly above the zero isotherm beginning at the 800-mb. level. By 1800 EST sleet had become mixed with the snow, indicating a continued increase in the above-freezing layer and also a deep cold layer remaining near the surface. At 2200 EST the advection of warm moist air had produced a layer of above freezing temperature nearly 5,000 feet thick, and the precipitation was entirely sleet. Sleet changed to freezing rain at 0300 EST with the 0400 EST sounding indicating a below-freezing surface layer less than 2,000 feet deep. By 1000 EST only the immediate surface level was below freezing, and within the next 2 hours temperatures rose above 0° C. The warm front passed the station at that time, changing the freezing rain to occasional rain.

For Washington, D. C. (fig. 6), a similar picture can be observed, but with minor exceptions. The most important deviation from the Albany pattern occurred at Washington at 2200 EST of January 9. This sounding indicated a definite change from the preceding and following observations, with an almost isothermal lapse rate from the surface to near 8,000 feet. This change may have been caused by some brief deviation in the upper air flow which resulted in a decrease of the warm air advection over Washington, thus allowing the snow falling through this inversion layer to cool the air to near the 0° C. isotherm. Wexler, Reed, and Honig [11] have stated that cooling of the air by melting snow will first cause an isothermal layer to form aloft at the freezing point. This cooling aloft creates an unstable lapse rate just below the layer of melting snow, so that cold air may be transported downward. It might be well to point out that during this period of the near-isothermal lapse rate the precipitation on the surface was considerably heavier than it had been earlier, and that the types of hydrometeors varied from freezing rain to sleet to snow.

The 1000 EST sounding of the 9th indicated a small layer with a superadiabatic lapse rate. It is opined that this is the result of the temperature element acting as a wet bulb and therefore failing to register ambient temperatures correctly for a short distance upon entering a layer of dry air after passing through a saturated layer.

On these series of soundings from Albany and Washington, two inversions will be noted. In the early stages the potential temperature of these surfaces was near 300° and 288° A., respectively. The lower isentropic surface was associated with the warm front on the polar frontal system, and the upper potential temperature has been considered as the boundary of the tropical air aloft.

Various studies have been made of the correlation between the occurrence of snow or rain and the upper-air thickness values. It has been observed by Lamb [12] that a 1000–500-mb. thickness value of 17,300 feet is critical over the United Kingdom as the determining point between snow or rain-type precipitation. The cutoff point is not sharply defined but higher thickness values usually produce rain. Over the United States, the Analysis Center finds the critical thickness value for stations near sea level close to the same value of 17,300 over the northwestern United States where air-mass contrasts are weak, but nearer 17,800 feet thickness over eastern United States where strong air mass contrasts produce considerably more stability. However, the cutoff of the precipitation type again is not sharply defined and will range slightly over 100 feet either side of the 17,800-feet thickness height. The 1000–500-mb. thickness values are entered to the top right of each of the plotted soundings in figures 5 and 6.

11. STABILITY CONSIDERATIONS

Air flowing over the Northeastern States during this period was stable. This was established by examination and comparisons of the thicknesses of various layers of the soundings throughout the region, following the procedure developed by Showalter [13]. Thickness values for the following layers were considered: 1000–500 mb. (henceforth referred to as Z_5); 1000–700 mb. (Z_7), and 1000–850 mb. ($Z_{8.5}$). Figure 7A, B, and C, shows the result of the investigation, charting values of Z_7 minus $Z_5/2$ and values of $Z_{8.5}$ minus $Z_7/2$ for January 7 to 10. Under standard lapse rate conditions the value for $(Z_7-Z_5/2)$ is 550 feet and that for $(Z_{8.5}-Z_7/2)$ is 345 feet. Computed $(Z_7-Z_5/2)$ values of less than 500 feet indicate greater stability, while values greater than 500 feet indicate an increasing trend toward instability. Values near 300 feet are about the maximum of stability and near 700 feet the maximum of instability.

Similar values can be arrived at for $(Z_{8.5}-Z_7/2)$, but in this case, disregarding the sign, the reverse of the above is true. The higher values make for greater stability and the smaller values instability. The range in values here is from near 250 feet to about 550 feet, with the border line between the stable and unstable air near 370 feet. However, this latter value is variable, dependent upon the thickness of the $Z_{8.5}$ layer.

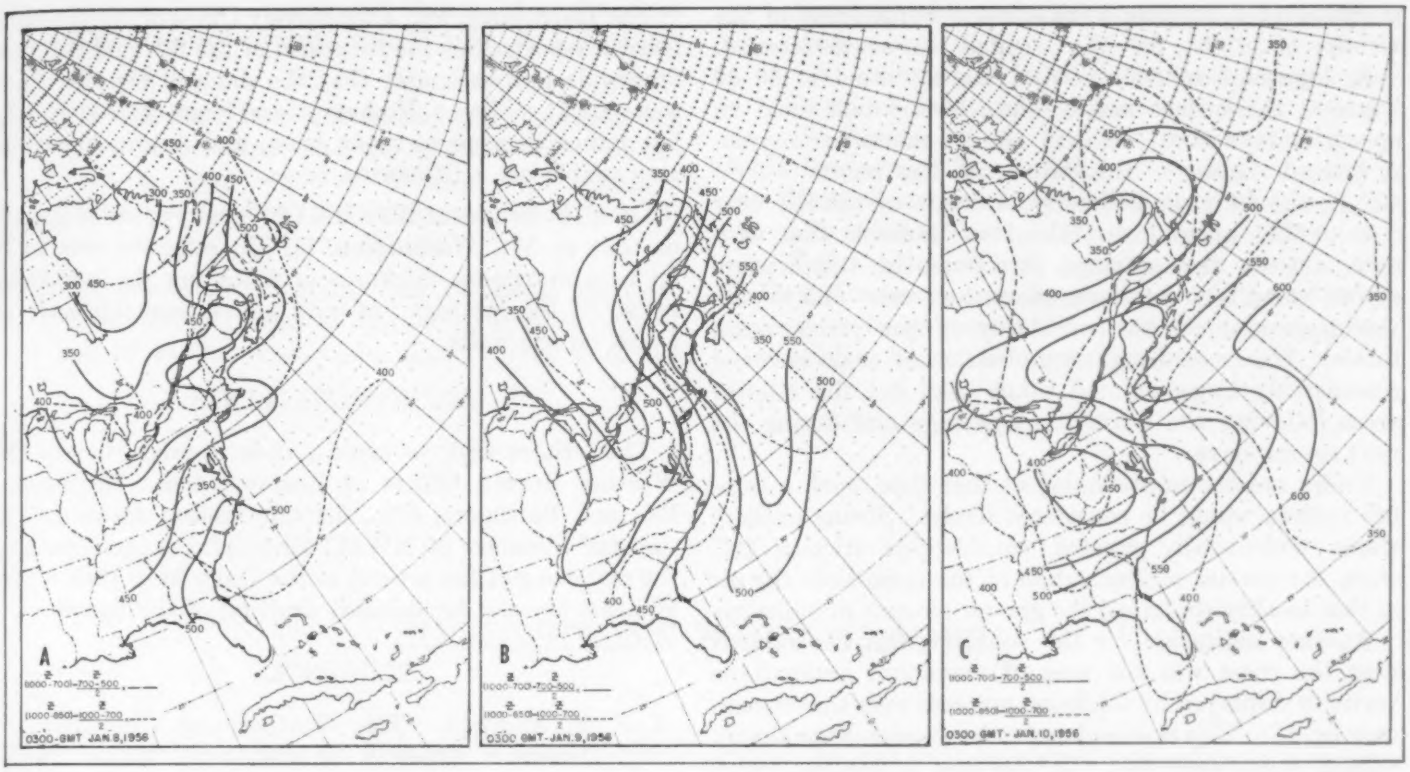


FIGURE 7.—Synoptic patterns of the stability and thickness relationships of two adjacent layers, January 8–10, 1956. (A) Composite chart for 0300 GMT, January 8. The 1000–700-mb. thickness value, minus the 1000–500-mb. thickness value divided by 2, is shown in tens of geopotential feet at 50-foot intervals by solid lines. Values below the 500-foot line indicate more stability and values above, a trend toward instability; the outside limits being near 300 and 700 feet, respectively. The dashed lines indicate the 1000–850-mb. thickness value, minus the 1000–700-mb. thickness value divided by 2, in tens of geopotential feet at intervals of 50 feet disregarding the sign, with departure from 370 feet indicating the reverse of the above, i. e., toward lower values would be increasing instability and higher values more stability. The limiting values for this layer are near 250 and 550 feet. (B) Composite chart for 0300 GMT, January 9, 1956. (C) Composite chart for 0300 GMT, January 10, 1956.

With these considerations in mind, it will be noted that the lines on the chart (fig. 7) indicate that the air from the surface to 500 mb. during this period was stable. However, there was some indication on the 10th that an area of instability was approaching the eastern seaboard in the vicinity of Boston. The value of the $(Z_{8.5}-Z_{7}/2)$ in that area was on the border of becoming conditionally unstable; this was also true of the value at the higher level. This instability in the eastern quadrants of the Low was borne out by ships at sea which reported thunderstorms on the 10th and 11th.

12. PRECIPITATION

PRE-WARM FRONT

Precipitation totals from the sleet and freezing rain were for the most part small and generally under one-quarter of an inch. That such was the case may be easily understood in a résumé of what has been discussed previously. As indicated by the temperature range of the soundings, the moisture content was low, ranging from 4 to 6 grams per kilogram. This would yield in stable air and on a normal warm front slope between 0.25 and 0.50 of an inch per day [14]. Along the coastal sections the duration of the freezing precipitation was short, thus a large accumulation of ice was not obtained. Over the interior the total accumulation for the most part was also small, even though the period of occurrence was of greater duration than in the coastal sections.

POST-WARM FRONT

As shown previously by the upper air temperature comparisons and by the soundings at Albany and Washington, it appeared that a modified form of tropical air was over the northeastern seaboard States. The inversion associated with this air had a potential temperature near 296° or 300° A. at the beginning of the storm period, and gradually lowered to near the 292° or 294° potential temperature surface after the warm front passage. A cross section on the 9th at 1500 GMT (not shown) placed the 292° potential temperature surface at the following heights: Buffalo 7,000 feet, Albany 5,500 feet, Portland 5,000 feet, and Sable Island 2,500 feet. This would yield an approximate isentropic slope of 1 mile in 800 miles. This potential temperature surface could be

extended to a possible weak surface warm front in the vicinity of 35° N., 58° W., oriented east-southeastward.

A slope of 1 to 800 with an easterly wind of 50–60 knots at the surface and 850-mb. level would yield an approximate vertical velocity of 0.03 meter per second. A vertical velocity of this value with dew points of 50°–55° F. should yield 24-hourly amounts of rainfall near 0.50 to 0.75 of an inch [14]. It is probable that with such a weak frontal slope this potential temperature surface was not uniform over the entire area. This in part would account for the variability of the precipitation totals. The coastal regions and adjacent areas received considerably larger rainfall totals than did the interior areas following the warm front passage and during the next several days.

Under conditions prevailing at that time, such a rainfall pattern would be considered likely. Strong onshore winds undoubtedly created considerable friction [15] along the coastal regions, while at the same time the air in that locality contained the greater amount of moisture.

Another indication for the continued heavier rainfall near the coast was the area of maximum anticyclonic vorticity displayed by thickness patterns over that region. (See fig. 1.) This is considered to be an area of maximum rainfall occurrence [16]. In this case it was not transported, but persisted over the coastal area from Boston to Portland throughout the period.

14. EFFECTS OF THE STORM

By far the most general effect of the ice storm was the delay and inconvenience inflicted on transportation facilities. The rash of minor traffic accidents at times exceeded one per minute in a few areas. Pedestrian accidents from falling on the slippery surfaces resulted in numerous broken bones, and bruises, as well as several deaths due to head injuries. Other deaths were ascribed to motor accidents. The glazing in Buffalo damaged power lines and trees. In several portions of New England the power companies had to produce heating on the high tension lines to thaw the accumulation of glaze. For short periods of 2 to 5 hours, it was necessary to discontinue bus service in certain communities, due to ice on roads and highways. Sanding and salting of the streets at times could not keep pace with the accumulation of glaze.

The long easterly fetch produced high tides from Delaware northward inflicting considerable damage on shore property and installations. At Atlantic City, N. J. tides of 4.5 feet above normal were recorded, and described as farther above normal than those of the 1954 and 1955 hurricanes. Philadelphia reported considerable damage from strong and gusty winds on the 10th. Blue Hill Observatory reported a peak gust of 65 m. p. h. on January 9, and a total rainfall that day of 3 inches. The arrival of the warm air and its persistence, in attendance with warm rains and melting of ice and snow, brought considerable flooding from ice-jammed streams.

The Harrisburg, Pa., Climatology Office of the Weather Bureau reports that the ice storm of the 8th and 9th was Statewide and estimates that the damage for the State was near $1 million. No heavy local losses were reported, but minor damage to roofs, trees, and automobiles was appreciable on a Statewide basis.

Dr. C. F. Brooks at Blue Hill Observatory, has informed us that at Mt. Washington, N. H., solid ice was built up to a maximum thickness of 6 feet on the northeast corner of the rampart but with an average thickness of 1 foot on the walls.

ACKNOWLEDGMENTS

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Articles are accepted for the Monthly Weather Review with the understanding that they have not been published or accepted for publication elsewhere.

Two copies of the *manuscript* should be submitted. All copy, including footnotes, references, tables, and legends for figures should be double spaced with margins of at least 1 inch on sides, top, and bottom. Some inked corrections are acceptable but pages with major changes should be retyped. The style of capitalization, abbreviation, etc., used in the Review is governed by the rules set down in the Government Printing Office Style Manual.

Tables should be typed each on a separate page, with a title provided. They should be numbered consecutively in arabic numerals.

In *equations* conventional symbols in accordance with the American Standards Association Letter Symbols for Meteorology should be used. If equations are written into the manuscript in longhand, dubious-looking symbols should be identified with a penciled note.

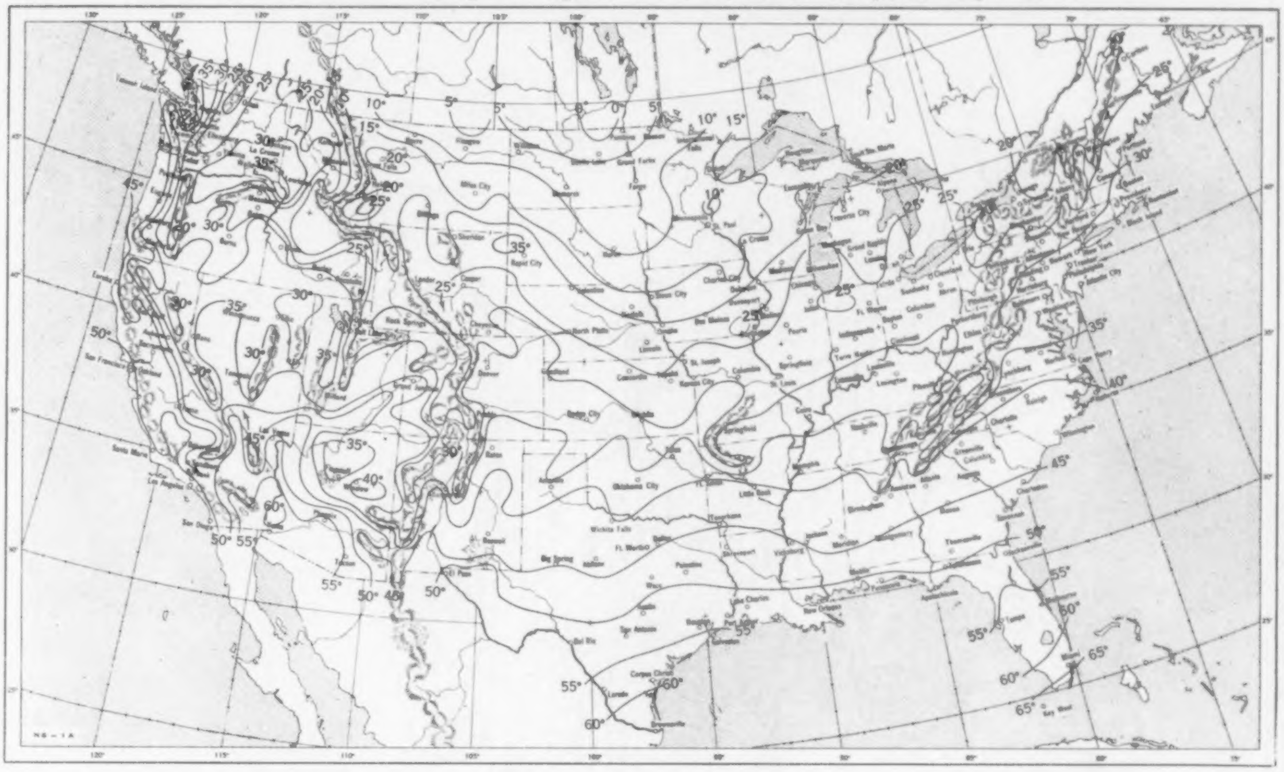
References should be listed on a separate sheet and numbered in the order in which they occur in the text; or if there are more than 10, in alphabetical order according to author. The listing should include author, title, source (if a magazine the volume, number, month, year, and complete page numbers; if a book the publisher, place of publication, date, and page numbers). If the referenced article is an independent publication, the author, title, publisher, place of publication, and date should be given. Within the text references should be indicated by arabic numbers in brackets to correspond to the numbered list.

Footnotes should be numbered consecutively in arabic numerals and indicated in the text by superscripts. Each should be typed at the bottom of the page on which the footnote reference occurs.

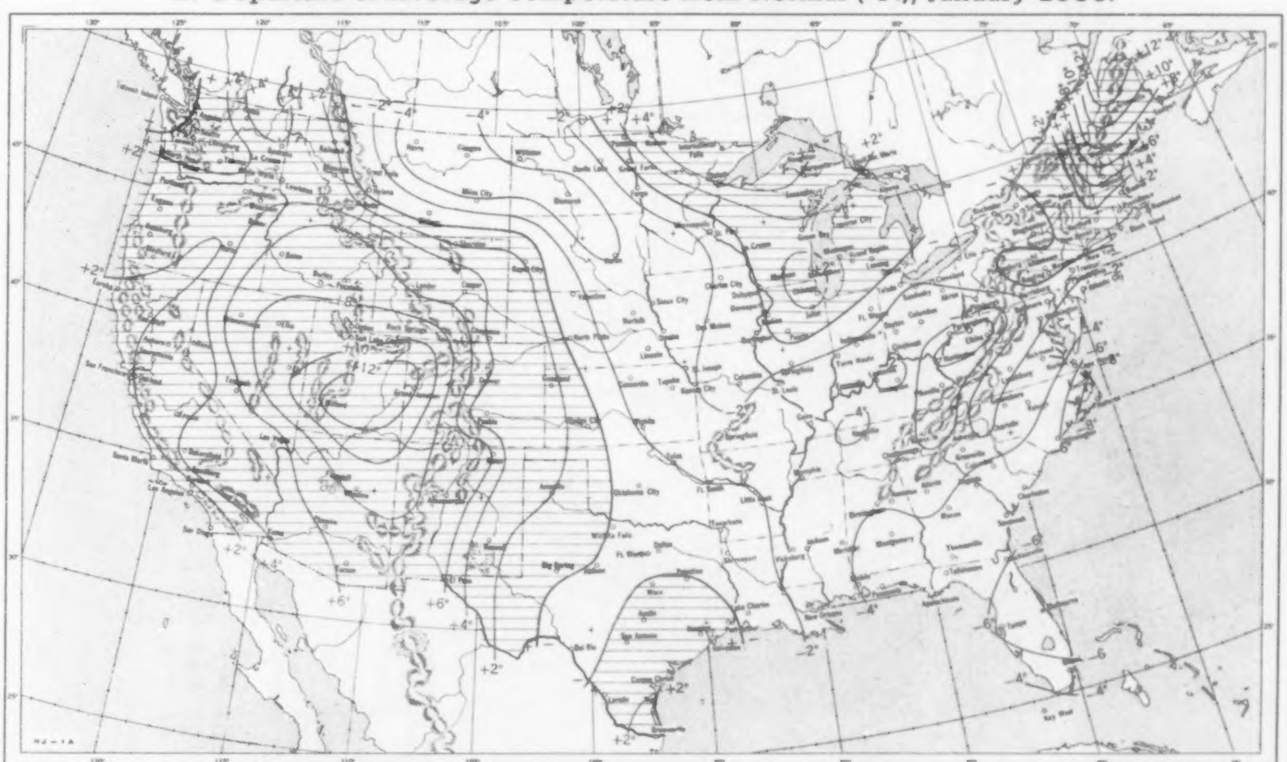
Illustrations. A list of legends for the illustrations should be typed (double spaced) on a separate sheet. Each illustration should be numbered in the margin or on the back outside the image area. To fit into the Review page, illustrations must take a reduction not to exceed 3½" x 9" (column size) or 7½" by 9" (page size). Map bases should show only political and continental boundaries and latitude and longitude lines, unless data are to be plotted, when station circles will also be needed. Usually the less unnecessary detail in the background the better will be the result from the standpoint of clear reproduction. Line drawings and graphs should also be uncluttered with fine background grids unless the graph demands very close reading. It is not necessary to submit finished drawings, as drafting work can be done at the time the paper is prepared for publication.

Photographs should be sharp and clear, with a glossy surface. Bear in mind that marks from paper clips or writing across the back will show up in the reproduction. Drawings and photographs should be protected with cardboard in mailing.

Chart I. A. Average Temperature (°F.) at Surface, January 1956.

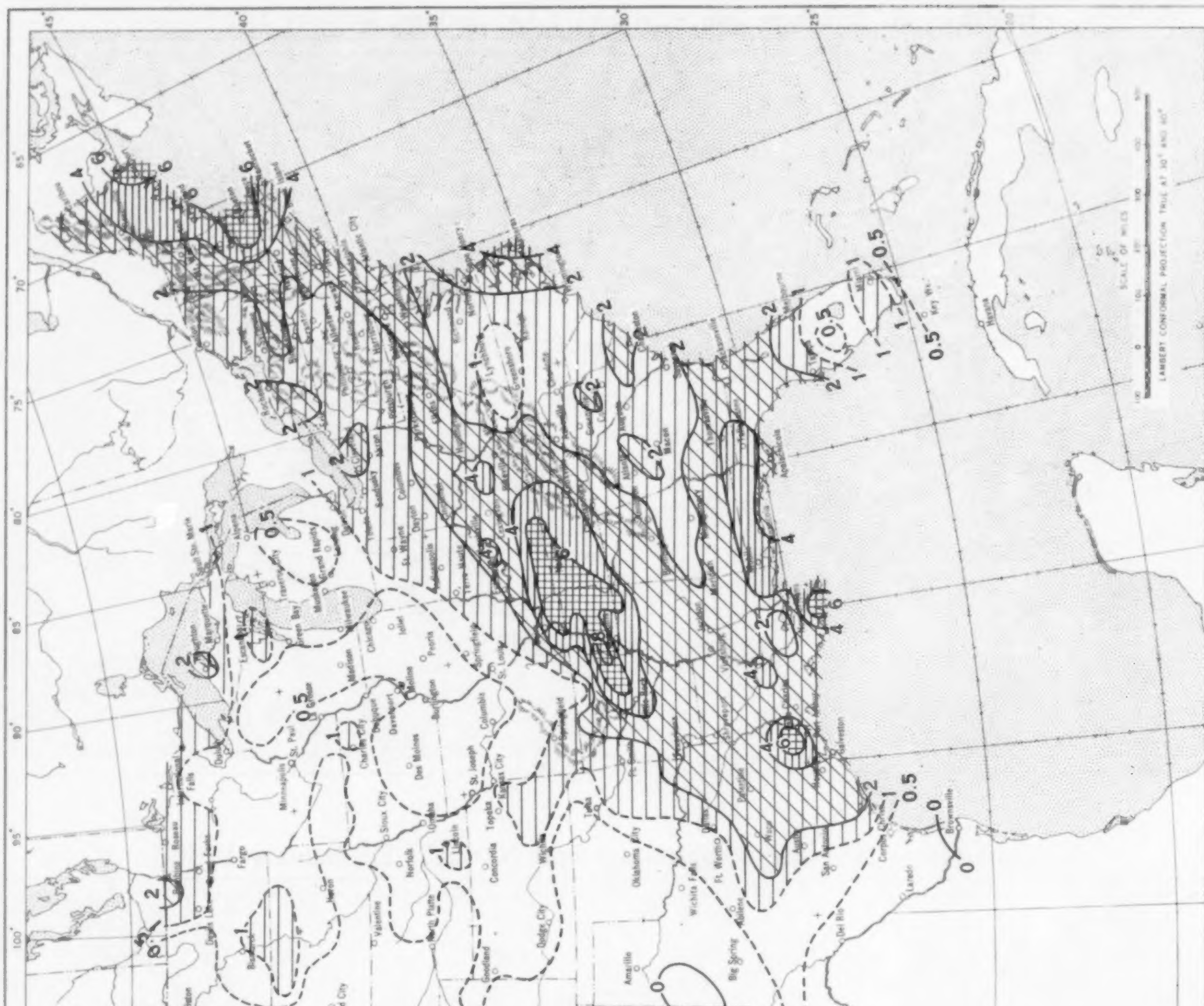


B. Departure of Average Temperature from Normal (°F.), January 1956.



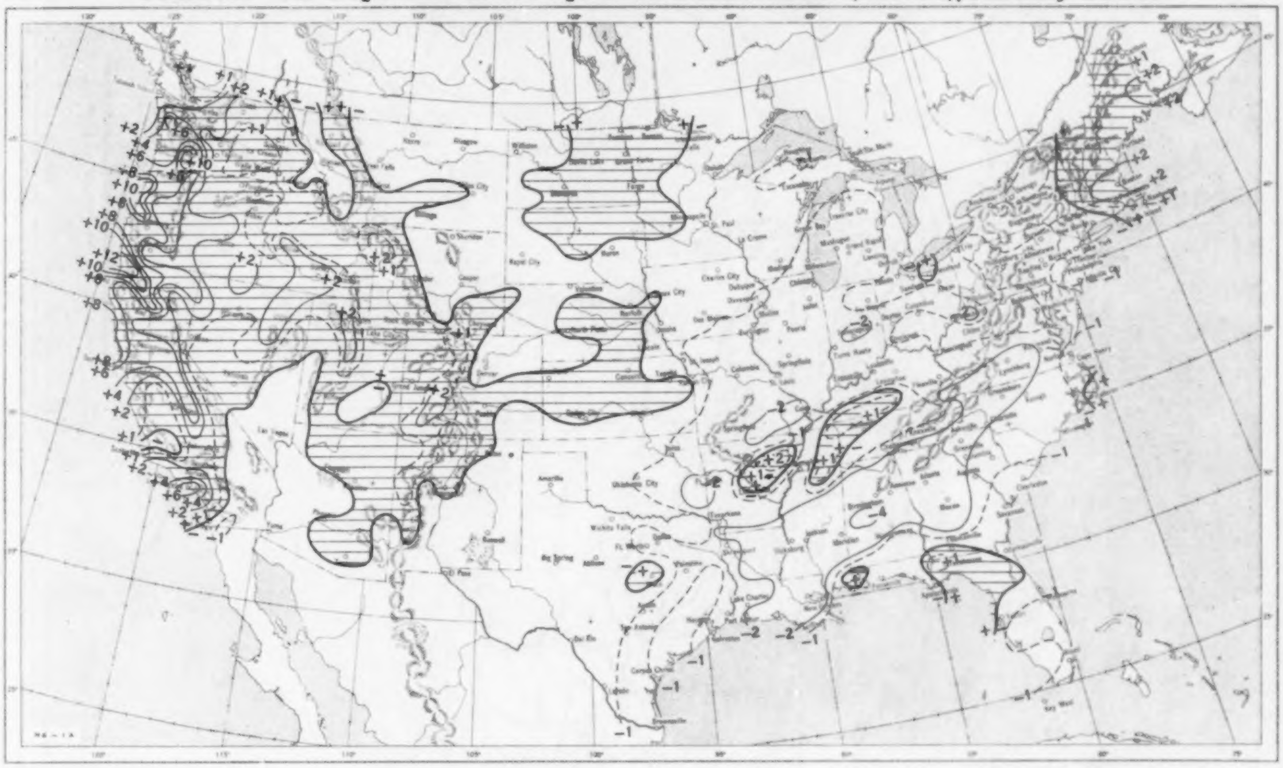
A. Based on reports from 800 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.
B. Normal average monthly temperatures are computed for Weather Bureau stations having at least 10 years of record.

Chart II. Total Precipitation (Inches), January 1956.

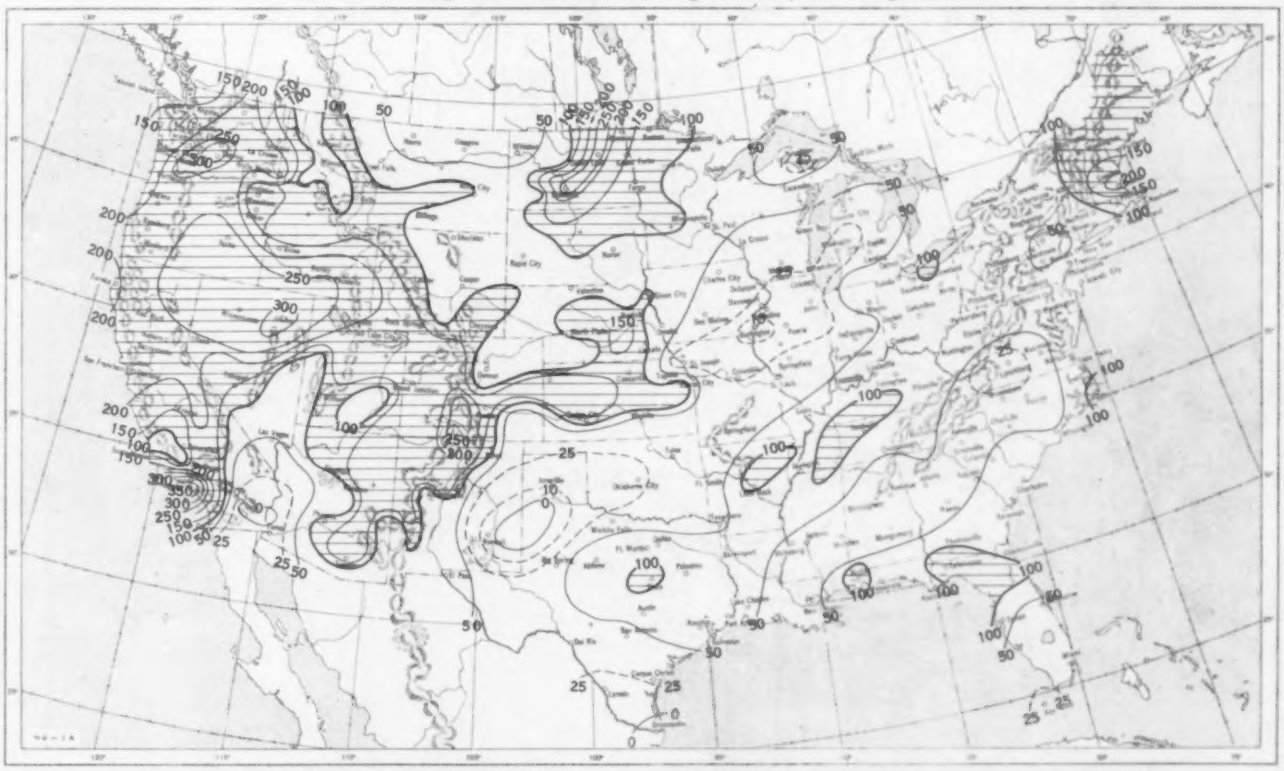


Based on daily precipitation records at 800 Weather Bureau and cooperative stations.

Chart III. A. Departure of Precipitation from Normal (Inches), January 1956.

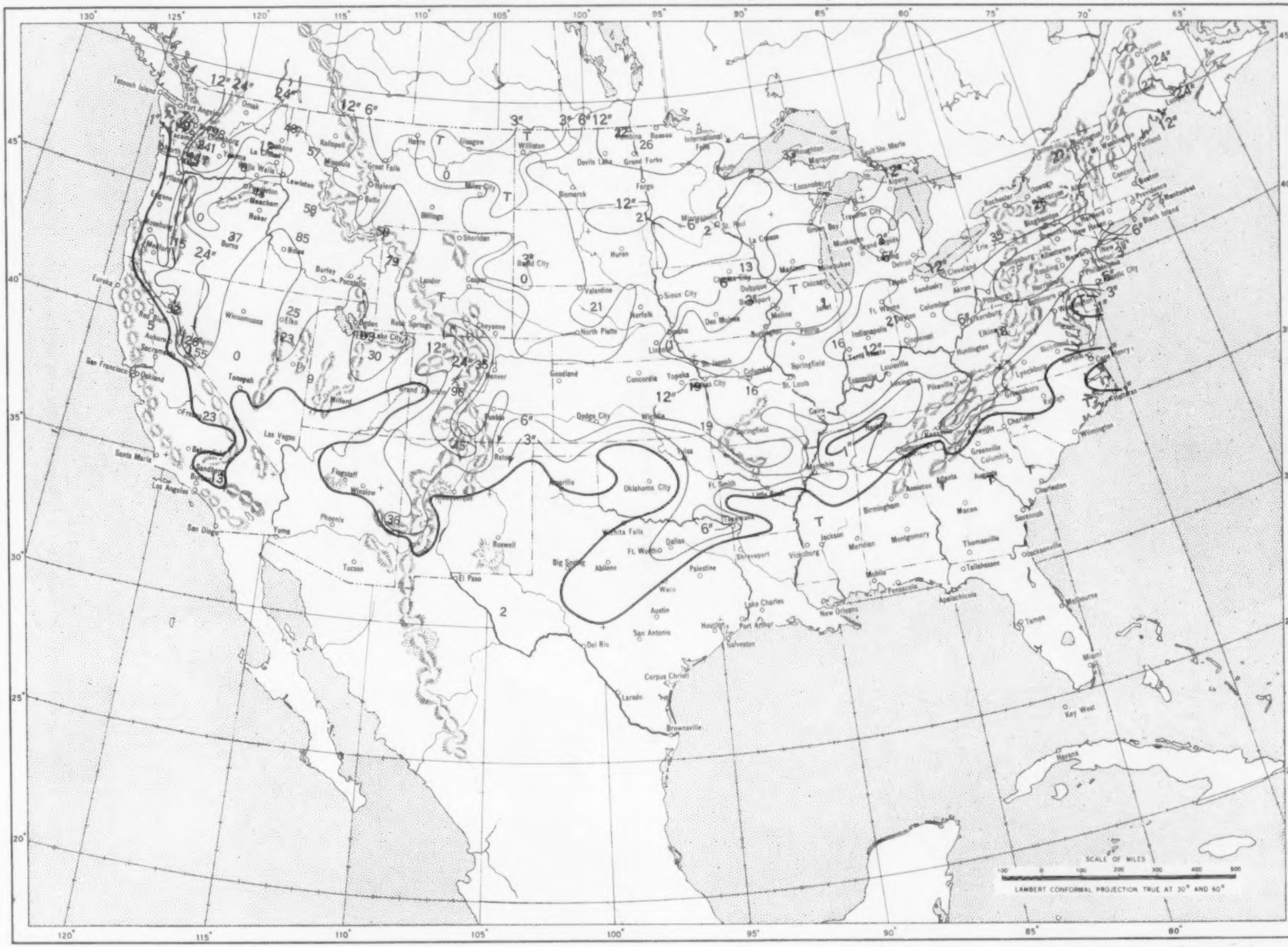


B. Percentage of Normal Precipitation, January 1956.



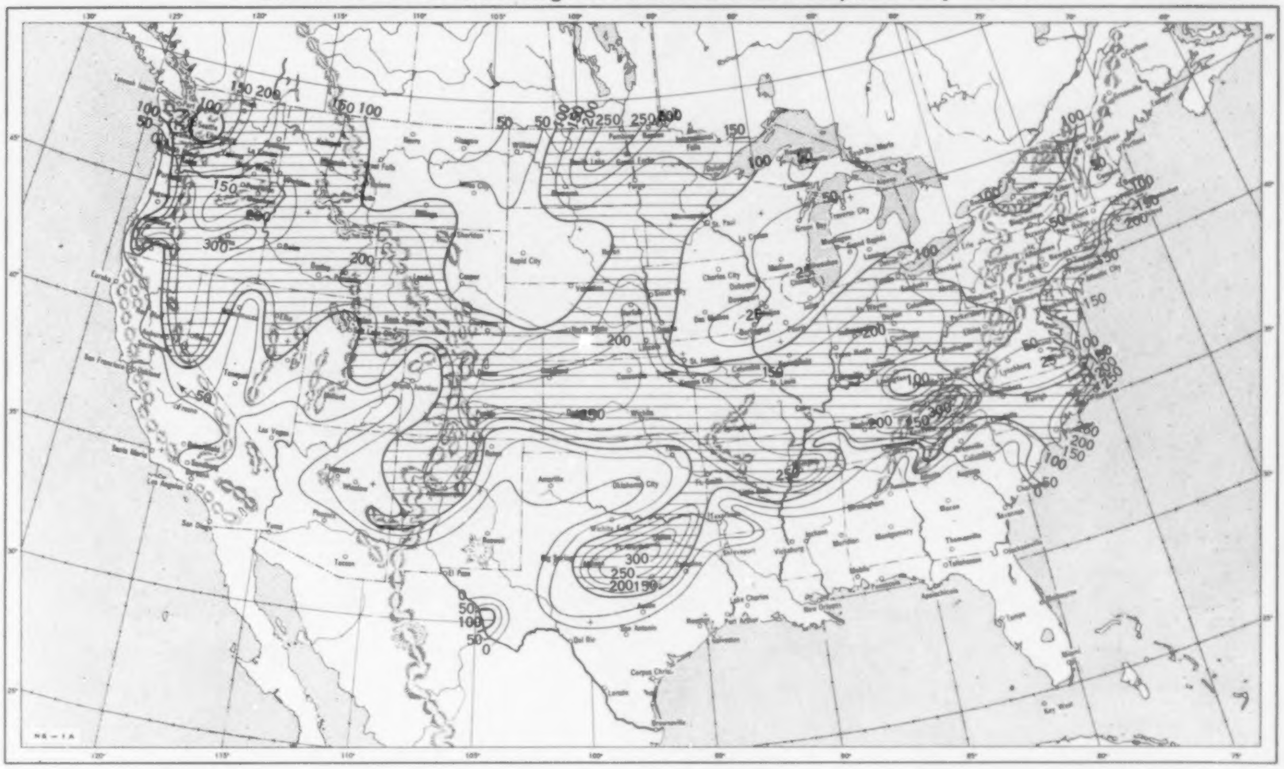
Normal monthly precipitation amounts are computed for stations having at least 10 years of record.

Chart IV. Total Snowfall (Inches), January 1956.

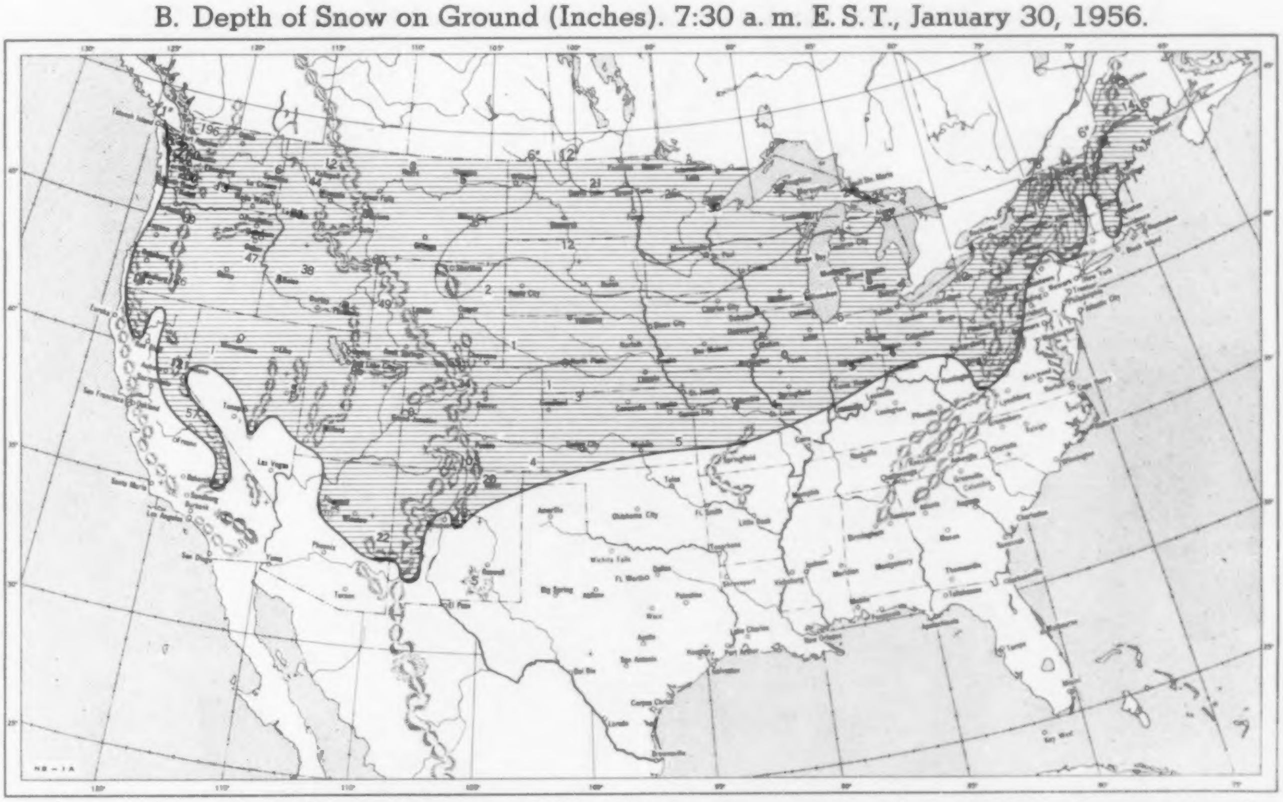


This is the total of unmelted snowfall recorded during the month at Weather Bureau and cooperative stations. This chart and Chart V are published only for the months of November through April although of course there is some snow at higher elevations, particularly in the far West, earlier and later in the year.

Chart V. A. Percentage of Normal Snowfall, January 1956.

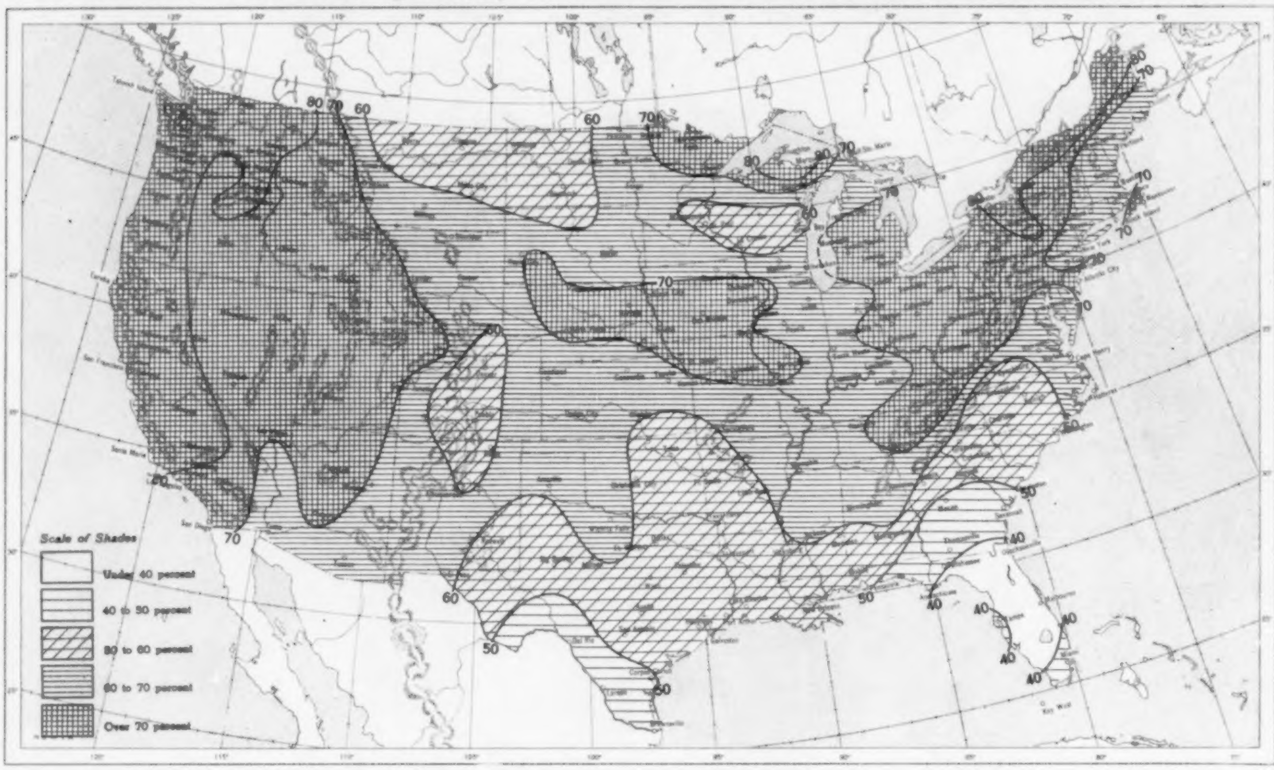


B. Depth of Snow on Ground (Inches). 7:30 a. m. E. S. T., January 30, 1956.

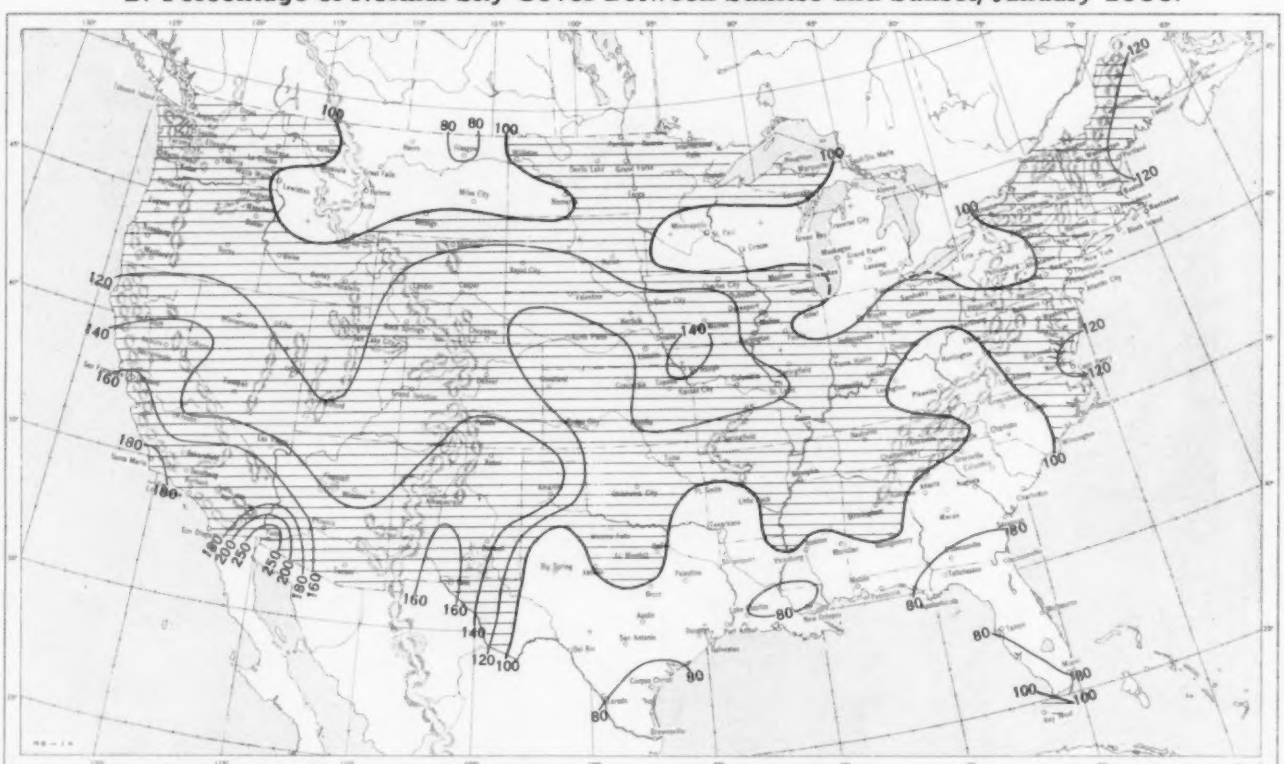


A. Amount of normal monthly snowfall is computed for Weather Bureau stations having at least 10 years of record.
B. Shows depth currently on ground at 7:30 a. m. E.S.T., of the Tuesday nearest the end of the month. It is based on reports from Weather Bureau and cooperative stations. Dashed line shows greatest southern extent of snowcover during month.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, January 1956.

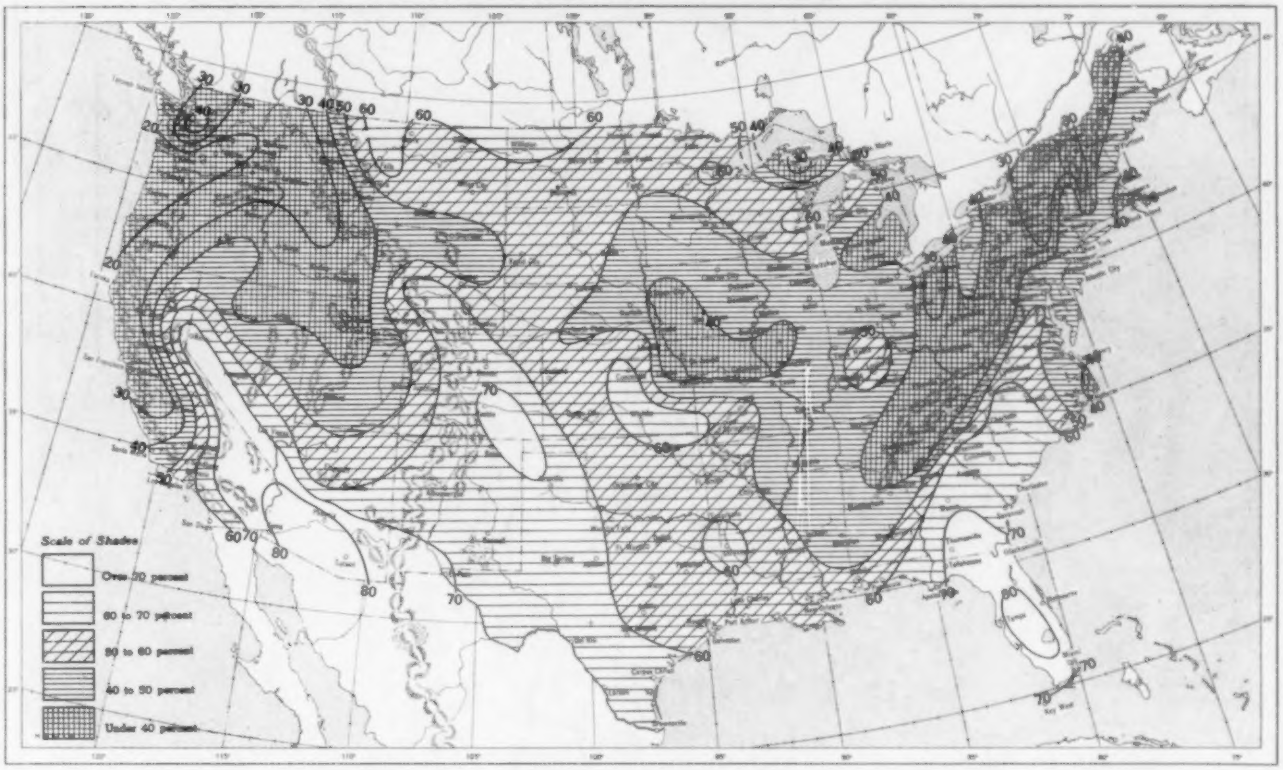


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, January 1956.

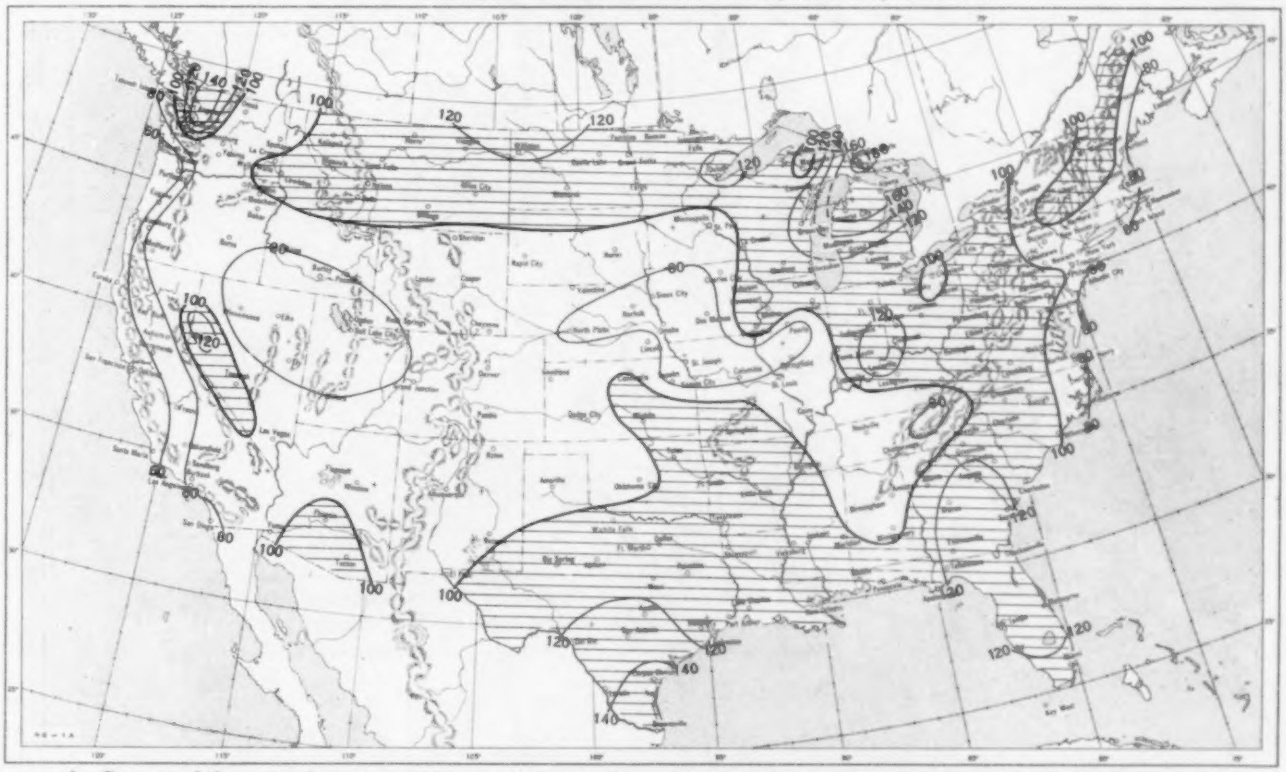


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, January 1956.



B. Percentage of Normal Sunshine, January 1956.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, January 1956. Inset: Percentage of Mean Daily Solar Radiation, January 1956. (Mean based on period 1951-55.)

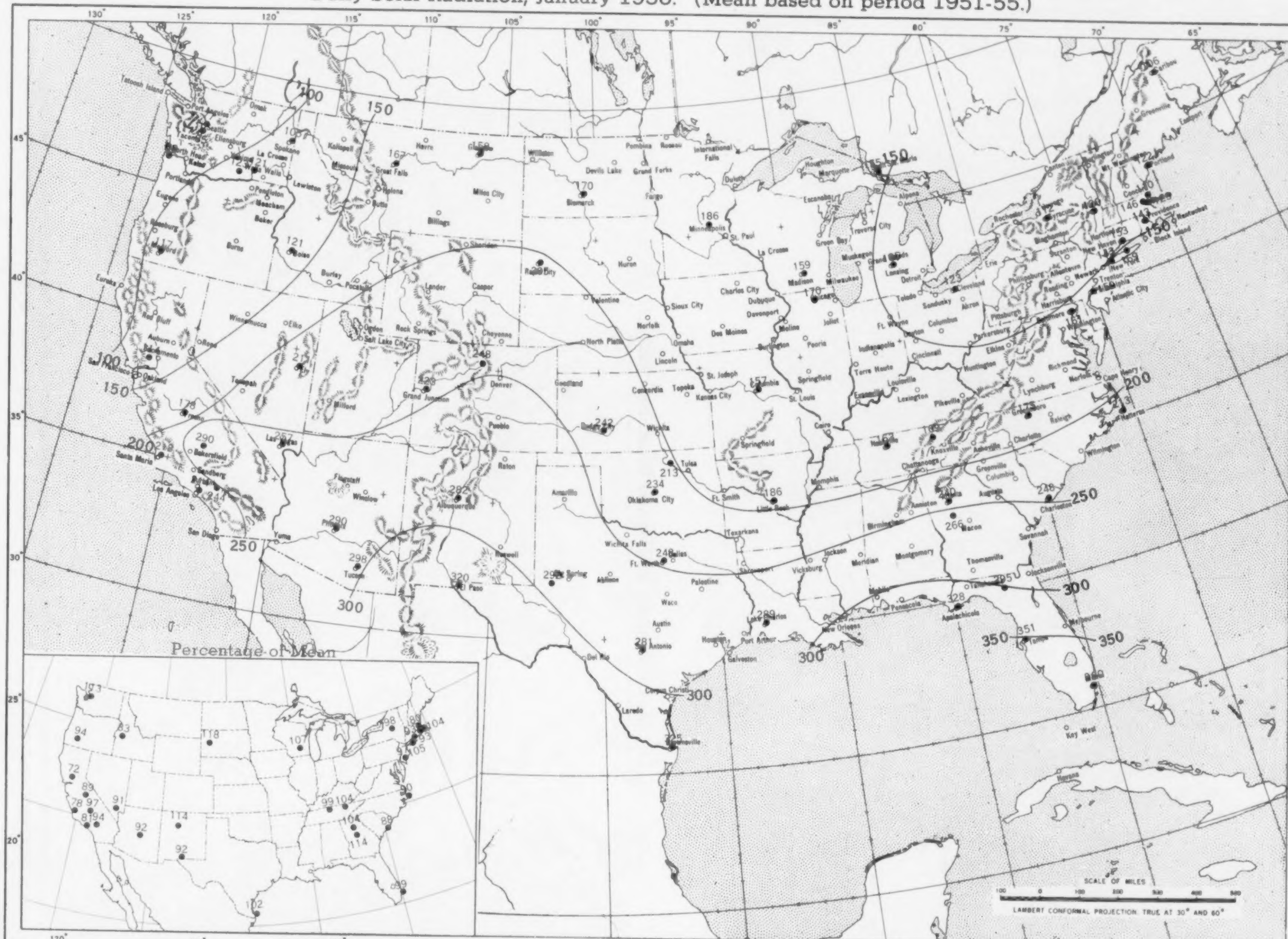
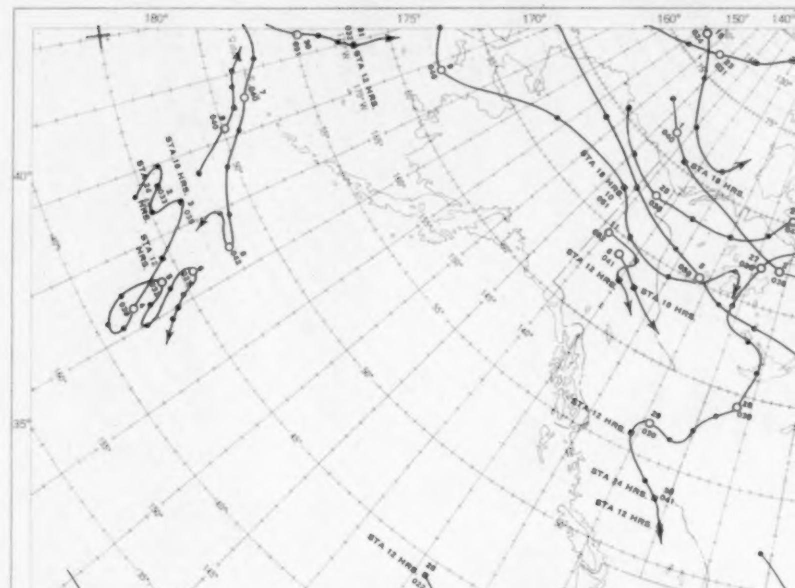


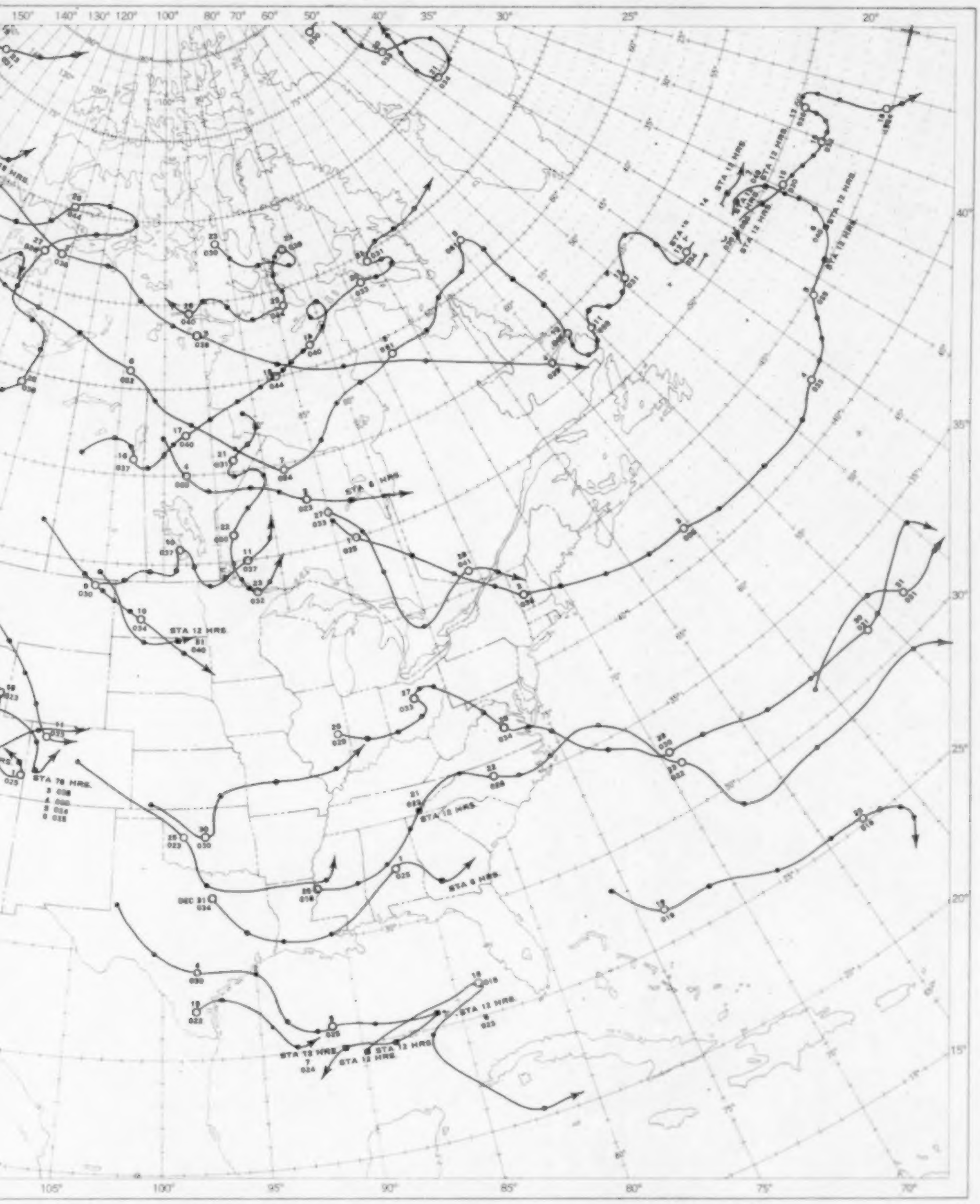
Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm. $^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown.

Chart IX. Tracks of Centers of Anti



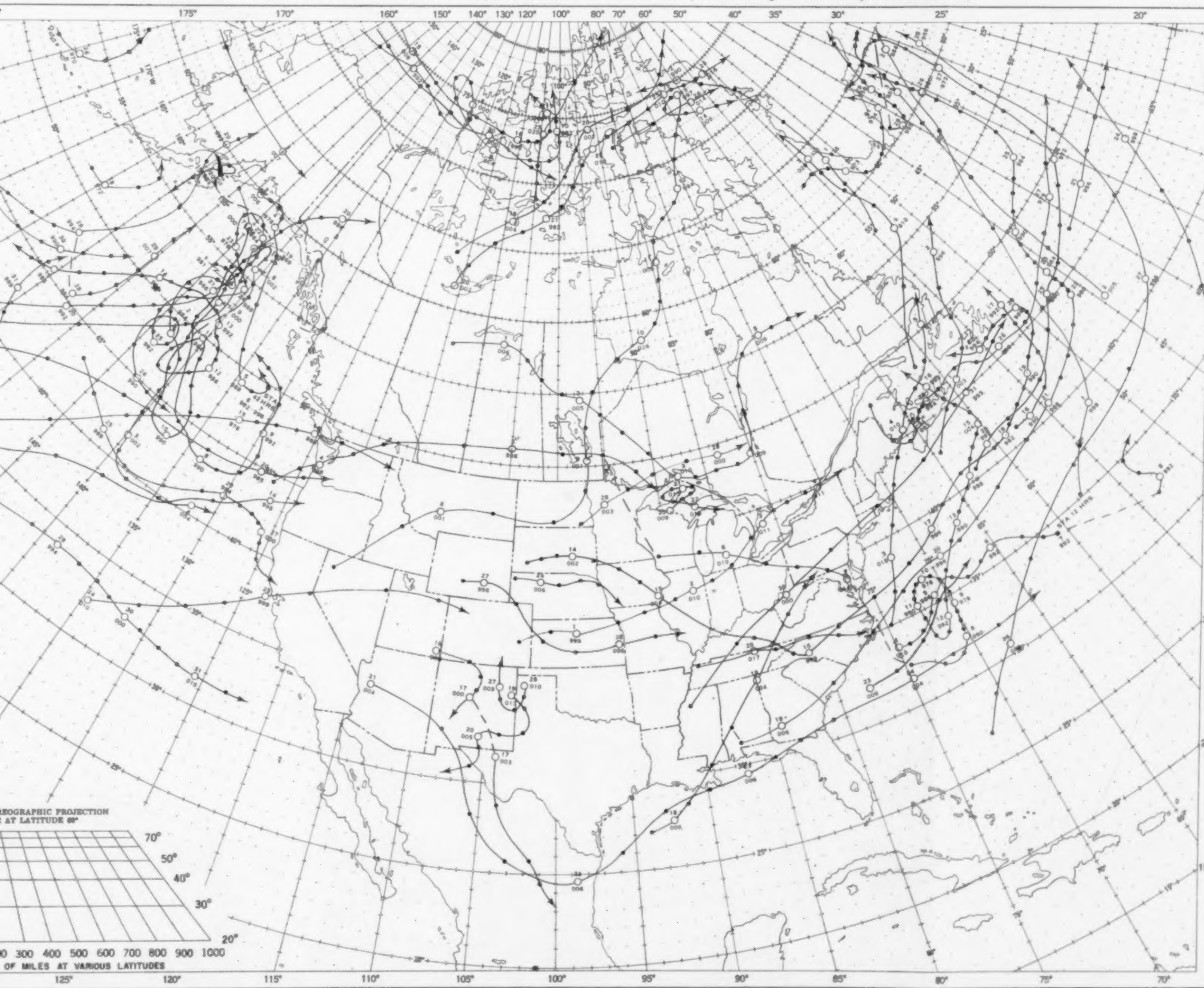
Circle indicates position of center at 7:30 a. m. E. S. T. Fi
Dots indicate intervening 6-hourly positions. Squares ind
indicates reformation at new position. Only those

of Anticyclones at Sea Level, January 1956. (Corrected)



S. T. Figure above circle indicates date, figure below, pressure to nearest millibar.
ares indicate position of stationary center for period shown. Dashed line in track
ly those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, January 1956. (Corrected)



Circle indicates position of center at 7:30 a. m. E. S. T. See Chart IX for explanation of symbols.

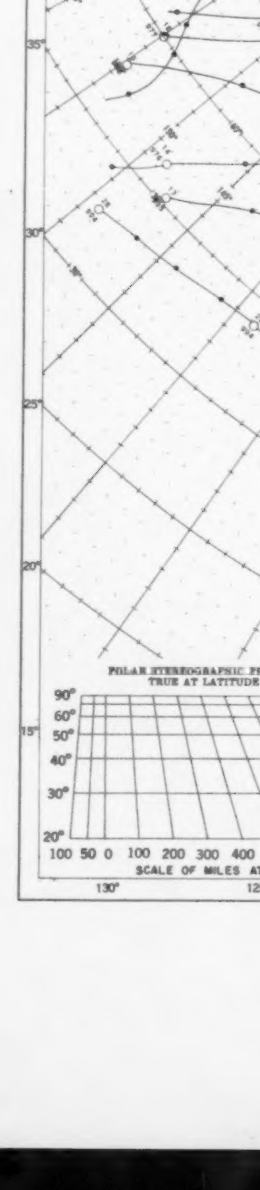
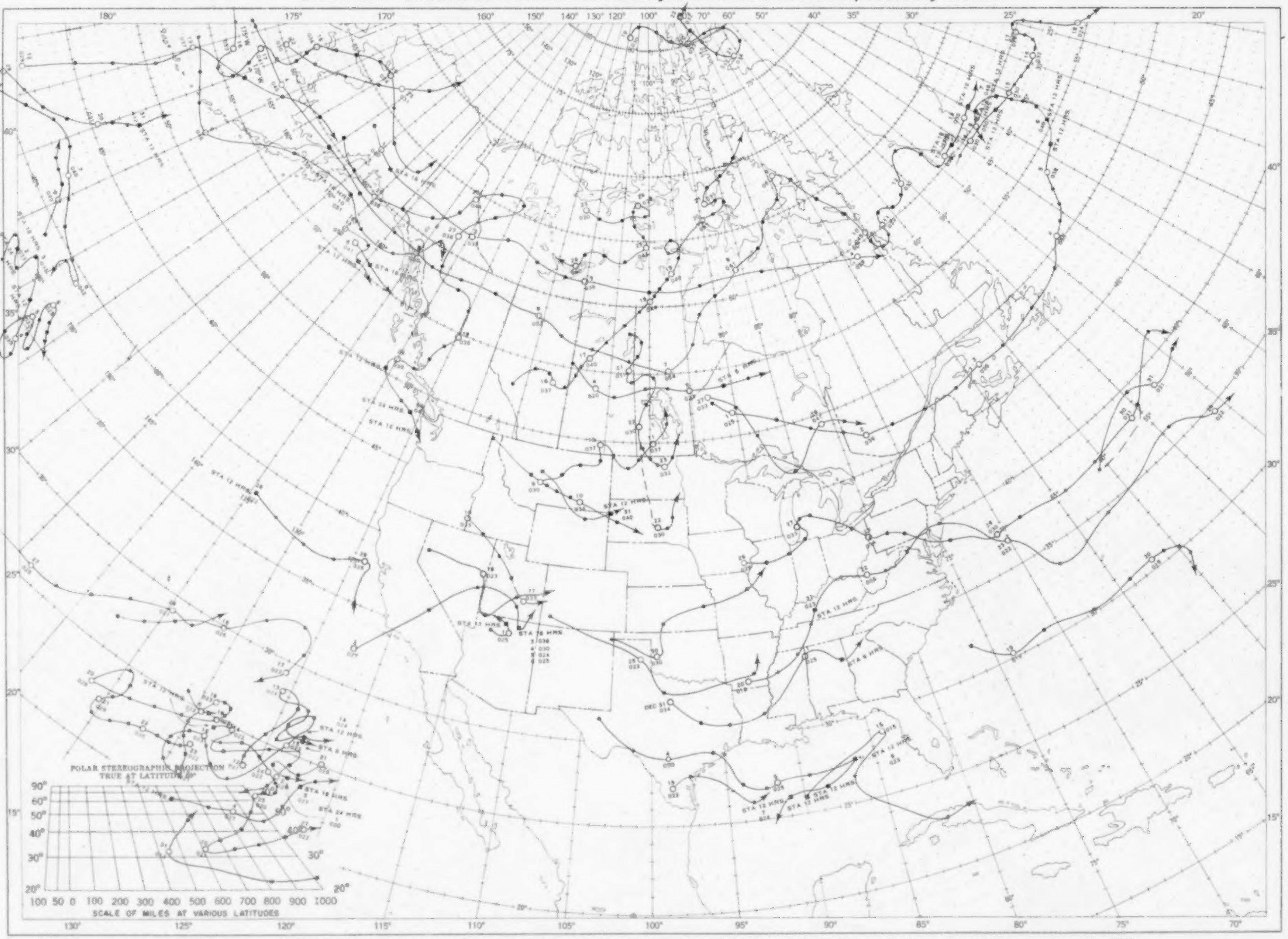
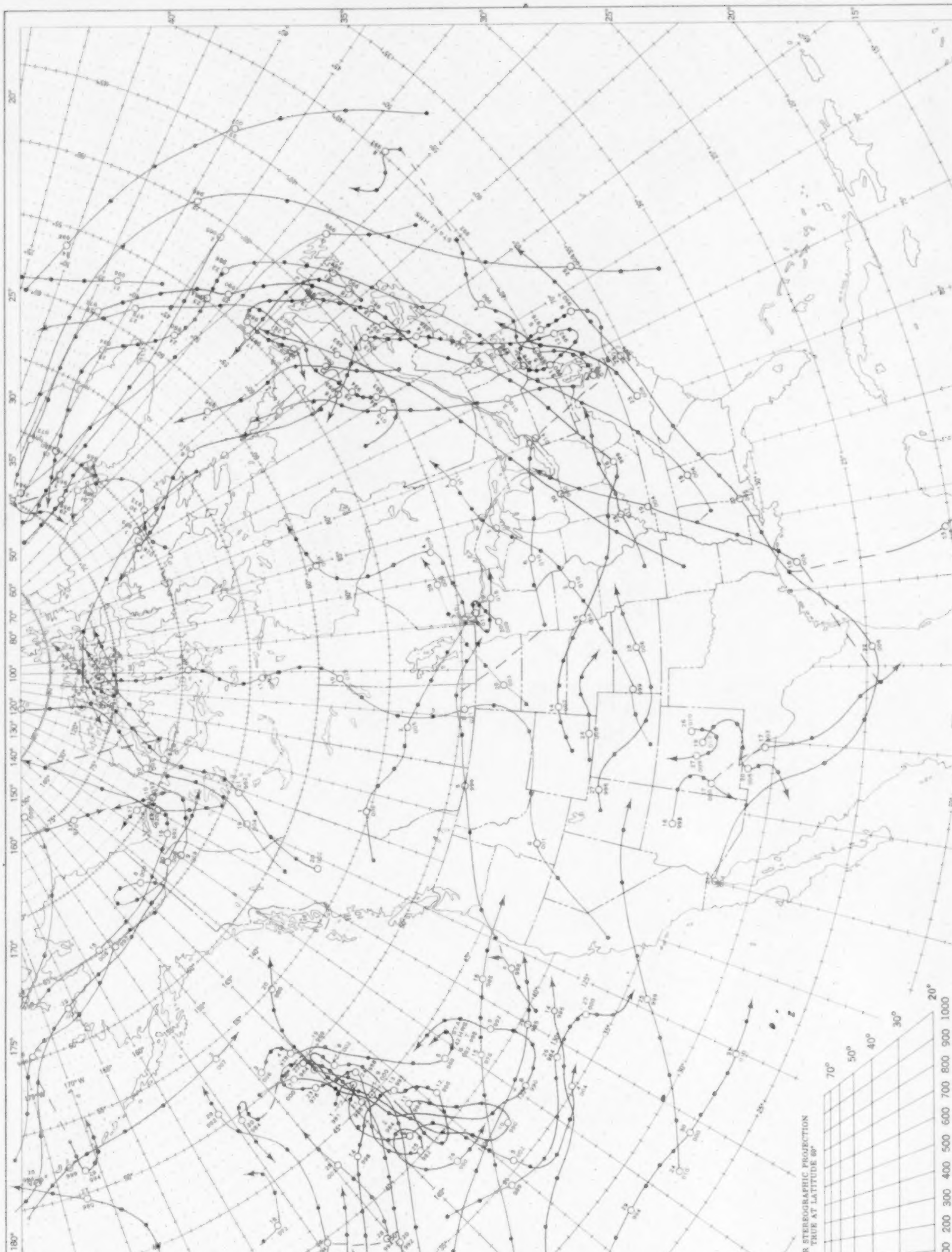

180°
40°
35°
30°
25°
20°
15°
POLAR STEREOGRAPHIC P
TRUE AT LATITUDE
90°
60°
50°
40°
30°
20°
100 50 0 100 200 300 400
SCALE OF MILES A
130°

Chart IX. Tracks of Centers of Anticyclones at Sea Level, January 1956.



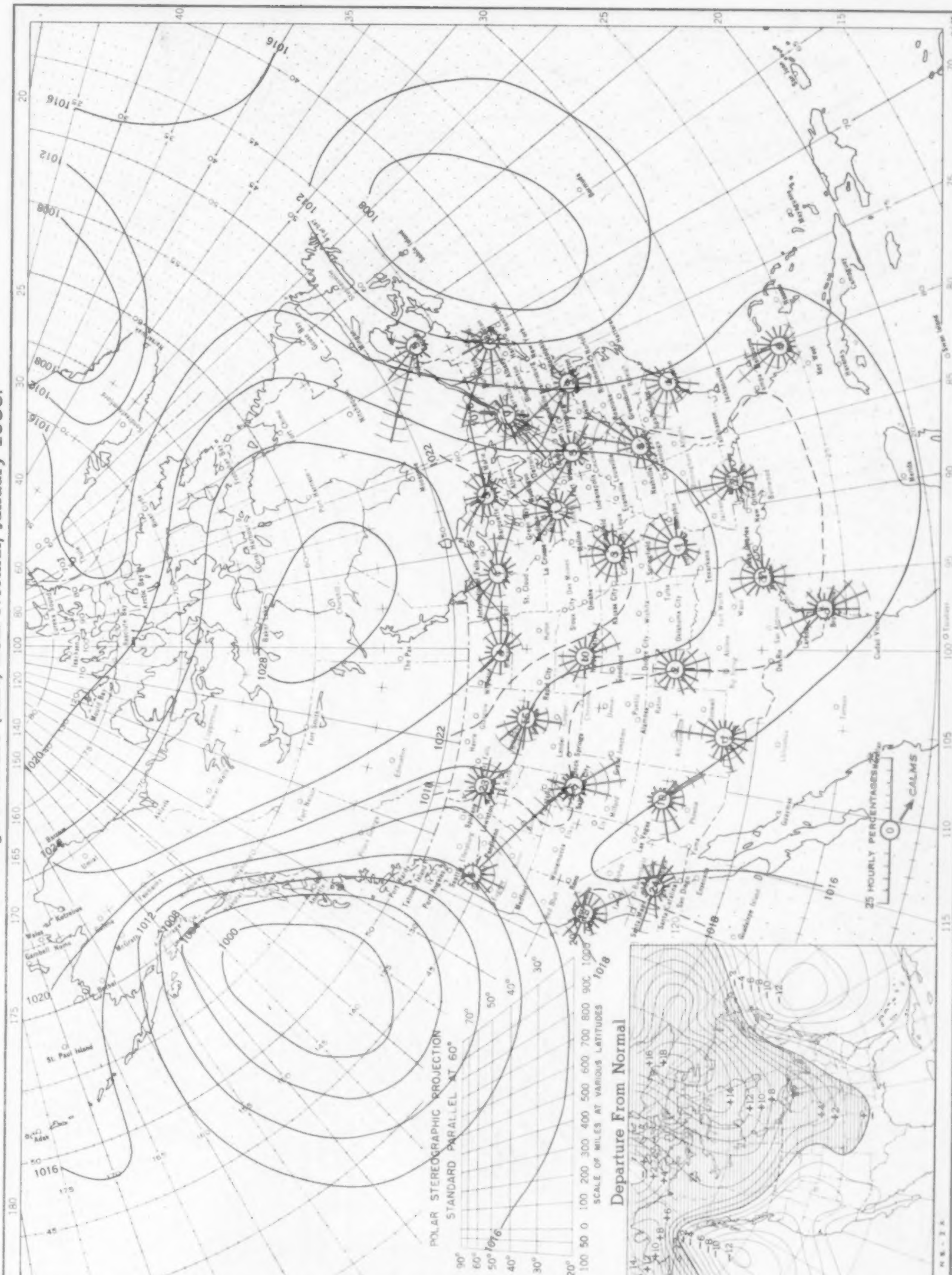
Circle indicates position of center at 7:30 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, January 1956.



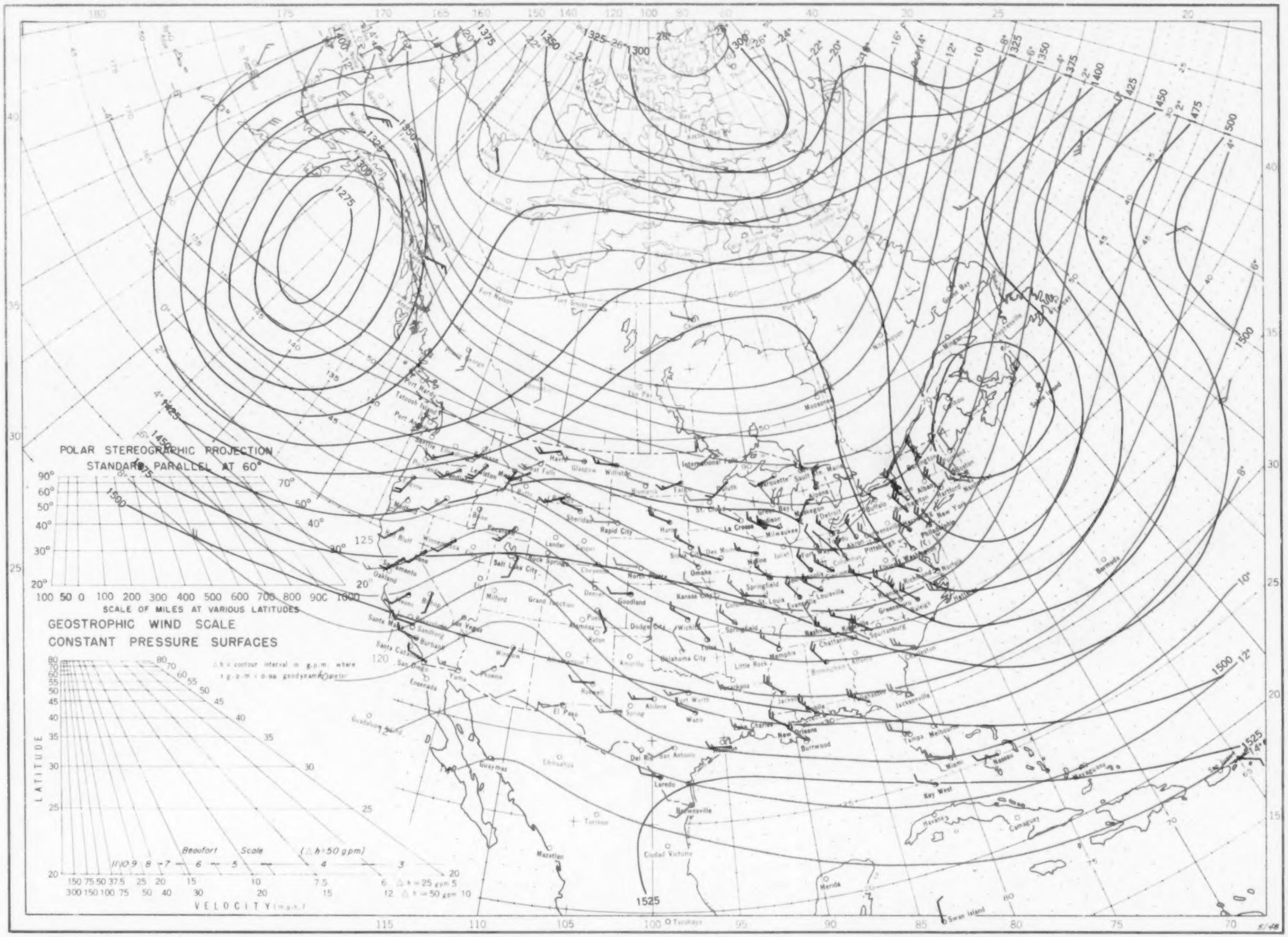
Circle indicates position of center at 7:30 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, January 1956. Inset: Departure of Average Pressure (mb.) from Normal, January 1956.



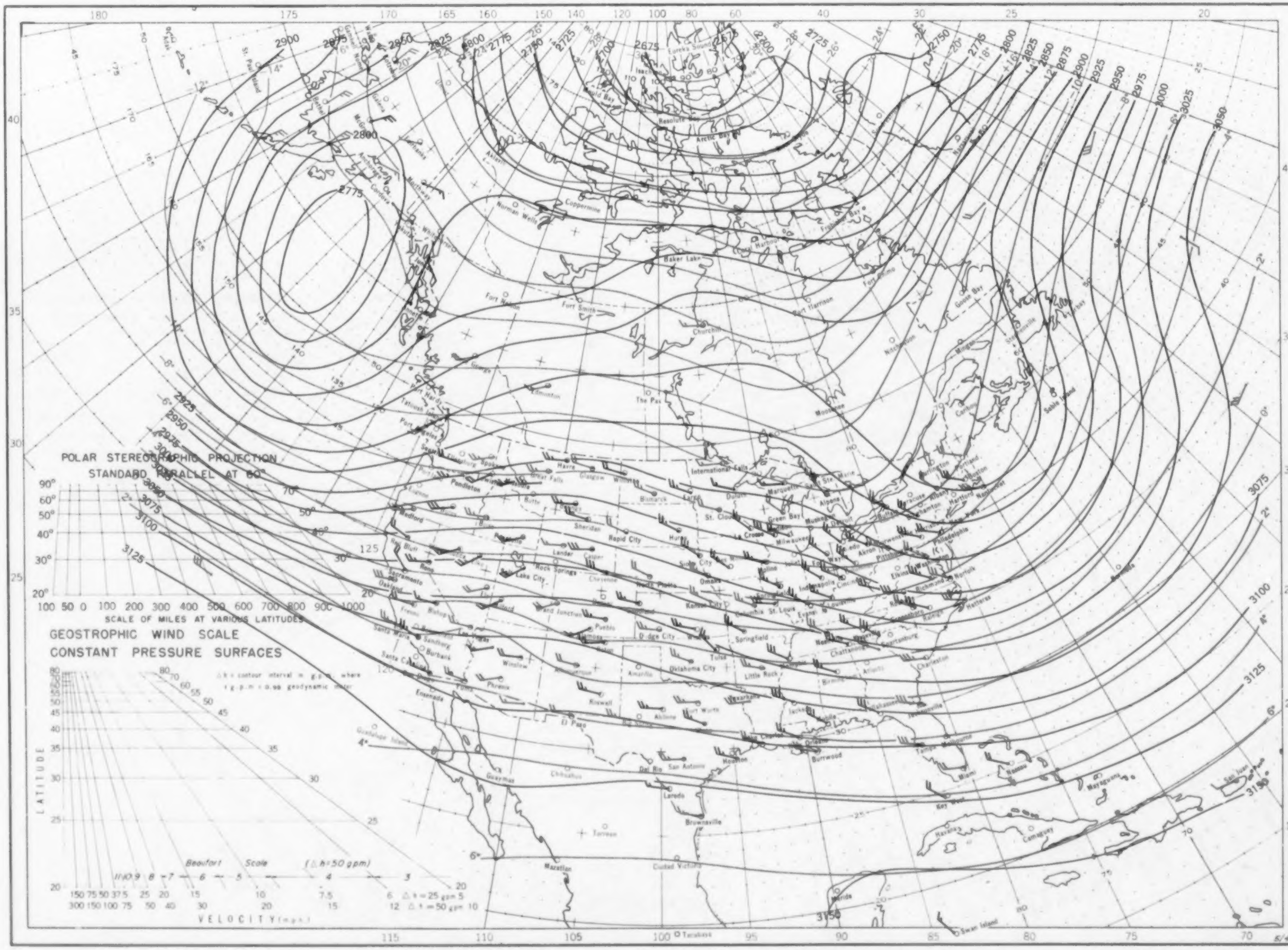
Average sea level pressures are obtained from the averages of the 7:30 a. m. and 7:30 p. m. E. S. T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 850-mb. Pressure Surface, Average Temperature in °C. at 850 mb., and Resultant Winds at 1500 Meters (m.s.l.), January 1956.



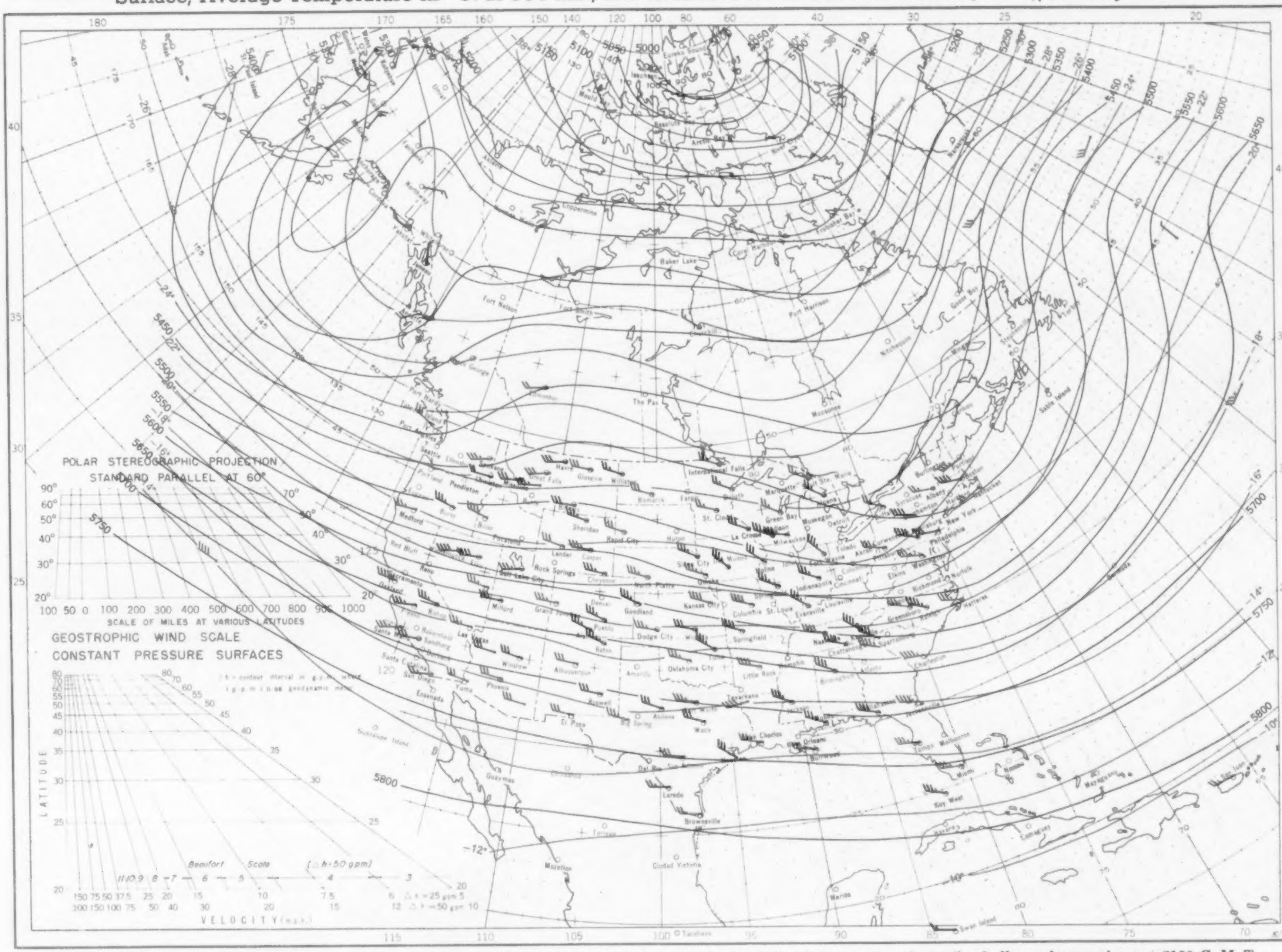
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.

Chart XIII. Average Dynamic Height in Geopotential Meters (1 g.p.m.. = 0.98 dynamic meters) of the 700-mb. Pressure Surface, Average Temperature in °C. at 700 mb., and Resultant Winds at 3000 Meters (m.s.l.), January 1956.



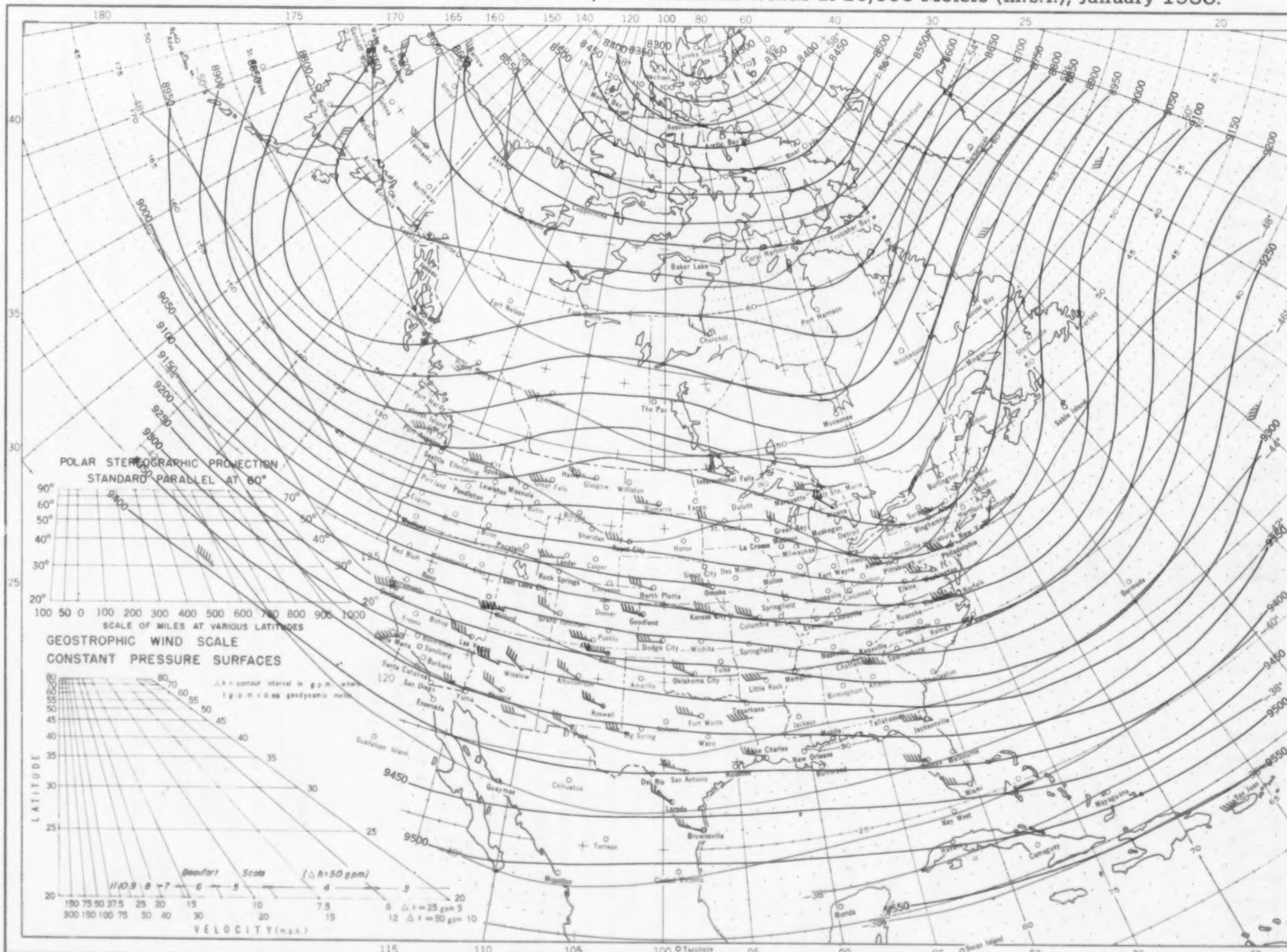
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.

Chart XIV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 500-mb. Pressure Surface, Average Temperature in °C. at 500 mb., and Resultant Winds at 5000 Meters (m.s.l.), January 1956.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.

Chart XV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 300-mb. Pressure Surface, Average Temperature in °C. at 300 mb., and Resultant Winds at 10,000 Meters (m.s.l.), January 1956.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.